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On the change of montly sea surface temperature in the Western North Pacific (I)

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Abstract: The time change of monthly sea surface temperature is closely correlated with the heat exchange between the sea and the atmosphere in the northern part of the Western North Pacific. This is supported by reasonable values of the depth of surface layer and of internal heat exchange which are obtained on simple assumptions. On the basis of the results, it is possible to represent the change of monthly sea surface temperature as a linear function of the heat exchange between the sea and the atmosphere for the area where the high correlation is found.

1. Introduction

The author examined the effects of advection and of diffusion on the ten-day or monthly mean sea surface temperature by using the correlation coefficients and spectra, and the persistence in the sea surface temperature values was evaluated (MORIYASU, 1967, 1968). In the study, the East China Sea was taken as a model case. Though the East China Sea is an appropriate area to the study, the results fell short of his expectations. This means that other factors, such as the heat exchange between the sea and the atmosphere and vertical motion and so on, are important to the variation of the sea surface temperature.

The heat exchange between the sea and the atmosphere is considered to be one of the important factors. Recently, WYRTKI studied the heat balance and the sea surface temperature of the North Pacific in the series of his works (1965 a, b, 1968) and obtained some useful results on the normal conditions. MASUZAWA (1952) and KOIZUMI (1956) discussed the heat exchange in the region of the ocean weather stations near Japan, and MOTOHASHI and UDA (1967) discussed the relation between the heat exchange and the monthly sea surface temperature. HISHIDA and NISHIYAMA (1969) evaluated the heat exchange in the Western North Pacific. These works contribute to the study on the heat budget

and serve as basic studies on the prediction of water temperature of the surface layer.

JACOB (1967) tried to predict monthly sea surface temperature, regarding advection as an important factor. In his study, the heat exchange was considered to have a secondary effect on the change of sea surface temperature. The present author, however, tries to examine the effect of the heat exchange on the variation of monthly sea surface temperature without regarding the effect of advection. The considered area is the Western North Pacific where climatological data are available.

2. Data and their processing

Climatological parameters in the Western North Pacific from 1962 to 1964 are compiled by Japan Meteorological Agency (1967 a, b, 1968). They are presented in the forms of monthly mean values of air temperature, sea surface temperature, dew-point temperature, cloudiness and wind speed for individual month. The processing is made for each quadrangle of 2 degrees in latitude and 5 degrees in longitude. Numbers of samples are dependent on quadrangles and months and, hence, each monthly mean value has various precision. The precision, however, is assumed to be constant for simplicity because of difficulty of the estimation. The region considered in this report is the sea north of the equator and from 120°E to 170°W, which consists of 377 quadrangles.

The amount of the change of monthly sea surface temperature (ΔT) for the ith month is defined as

$$\Delta T_i=\frac{1}{2}(T_{i+1}-T_{i-1}).$$

T_i means the monthly sea surface temperature for the ith month. The amount of the heat exchange (Q) between the sea and the atmosphere for the month is given by

$$Q=Q_s-(Q_b+Q_h+Q_e),$$

where Q_s is the incoming short-wave radiation, Q_b the effective back radiation, and Q_h and Q_e are the heat energy supplied from the sea to the atmosphere in sensible and latent forms, respectively. Those elements are not measured directly but are estimated by empirical formulas. Though, hereto, many formulas have been presented, it is not examined which formula is most appropriate to this study, but the following ordinary formulas are taken.

$$Q_s=Q_{s0}(1-0.71C)(1-r)$$

$$Q_b = s\sigma\theta_0^4(0.39-0.05\sqrt{e_a})(1-kC^2)+4s\sigma\theta_0^3(\theta_0-\theta_a)$$

$$Q_h = \rho c_p c_D(\theta_0-\theta_a)w$$

$$Q_e = \rho c_D L(q_w-q_a)w$$

Q_{s0} is the value of Q_s at the clear sky and its average monthly values are estimated as a function of latitude and month, and the albedo of the sea surface r is given as a function of latitude (BUDYKO, 1956). k is a parameter increasing with the latitude from 0.50 at the equator to 0.76 at 60° of latitude. Meanings of other notations are as follows:

- s : the ratio of the radiation of the sea surface to that of a black body (0.9 is used)
- σ : the Stefan-Boltzmann constant
- ρ : the density of the air
- c_p : the specific heat of the air at constant pressure
- c_D : the drag coefficient (1.55×10^{-3} is used)
- L : the latent heat of evaporation
- θ_0 : the absolute temperature of the surface water
- θ_a : the absolute temperature of the air
- C : the fractional cloudiness
- w : the wind speed
- e_a : the vapour pressure of the air in millibars
- q_w : the specific humidity at the sea surface
- q_a : the specific humidity of the air

θ_0, θ_a, C and w are given in the climatological table, and e_a, q_w and q_a are computed from air temperature, sea surface temperature and dew-point temperature in the table.

The processing is carried out for each quadrangle and each month. Thus, up to 34 pairs of ΔT and Q are obtained for a quadrangle, but the numbers of pairs are small in the southern and the northern parts of the region.

3. The change of the monthly sea surface temperature and the heat exchange

Fig. 1 shows the change of monthly sea surface temperature and the amount of the heat exchange for some quadrangles. The characteristic features of their annual variation are sensitive to the location, especially the latitude of the quadrangle, as pointed out by WYRTKI, and HISHIDA and NISHIYAMA. The amplitude of annual variation is larger in the northern part, while the variation is not distinguished in the southern part. The amount of the heat exchange is

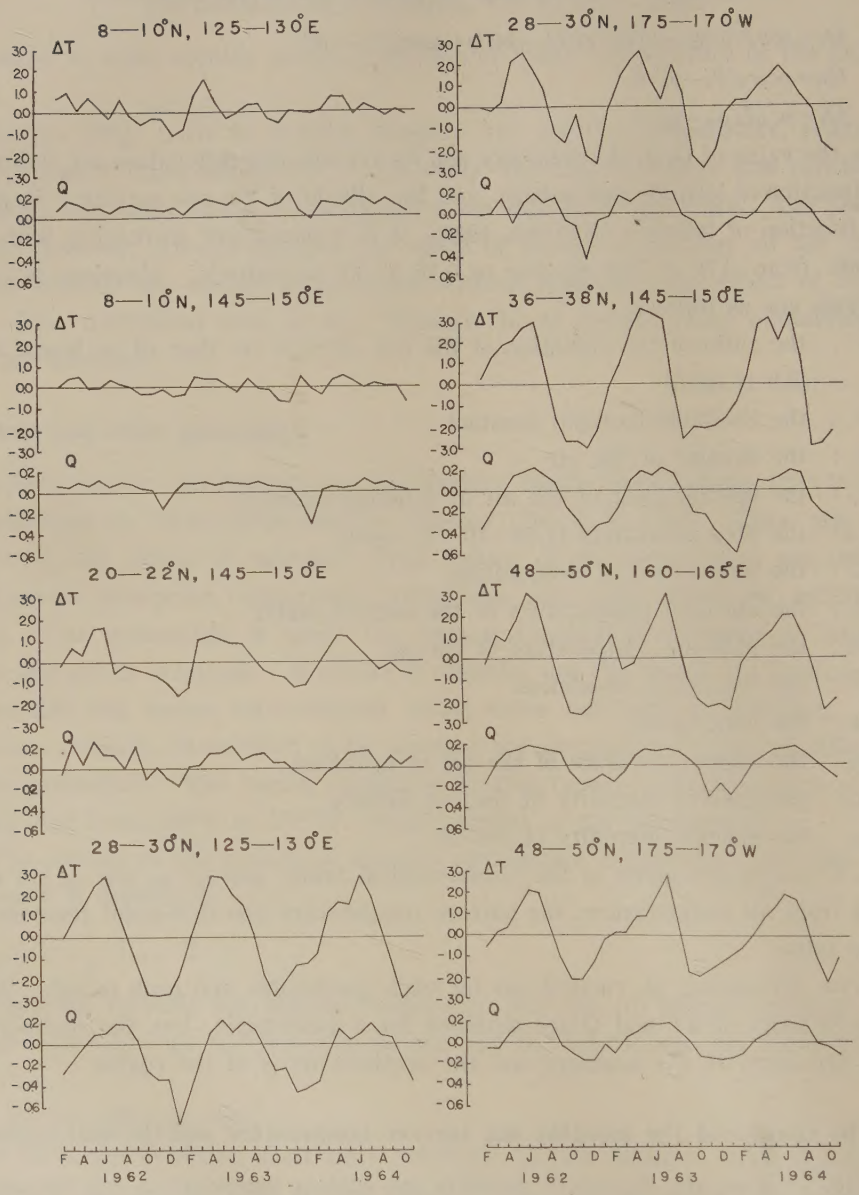

Fig. 1. The change of monthly sea surface temperature (ΔT) in °C/month and the heat exchange between the sea and the atmosphere (Q) in ly/min in some selected quadrangles.

positive almost all months in the southern part, while negative in the Kuroshio region of the northern part. The sea obtains the heat energy from the atmos-

phere in the southern part and the energy is carried northward by the Kuroshio. In the Kuroshio region of the middle latitudes, the heat energy is given to the atmosphere.

Except for the southern part, the phase of the change of monthly sea surface temperature is in good agreement with the phase of the amount of heat exchange. This feature suggests close relationships between them. Then, correlation coefficients are computed for each quadrangle. The number of samples to a quadrangle is at most 34 and computation is not carried out for quadrangles having less than 10 samples. The significant point at 95% level is 0.338 for 34 samples and 0.632 for 10 samples. Fig. 2 shows the distribution of the correlation coefficients.

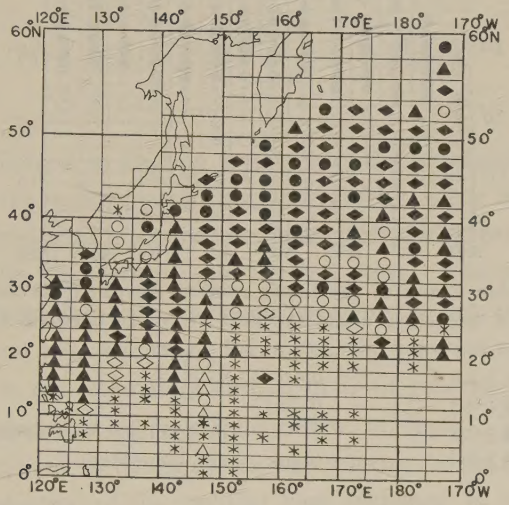

Fig. 2. Correlation coefficients between the change of monthly sea surface temperature (ΔT) and the heat exchange between the sea and the atmosphere (Q).

●≧0.8	0.5>◇≧0.4
0.8>◆≧0.7	0.4>△≧0.3
0.7>▲≧0.6	* insignificant at 95% level
0.6>○≧0.5	blank not computed

North of 26°N, insignificant values of correlation coefficients are not found and almost all values are greater than 0.6. South of this latitude, the correlation is not so high in many quadrangles, but values above 0.6 are found in the western part where the Kuroshio flows.

In this computation, the seasonal variation is not removed and the relation-

ship is close where the seasonal variation is marked. Therefore, it is referred that the basic seasonal variation of sea surface temperature is determined by the amount of the heat exchange for the region where the correlation coefficient is high. It is useful to represent ΔT as a linear function of Q for quadrangles where the correlation coefficient is significant. Fig. 3 shows the probable error

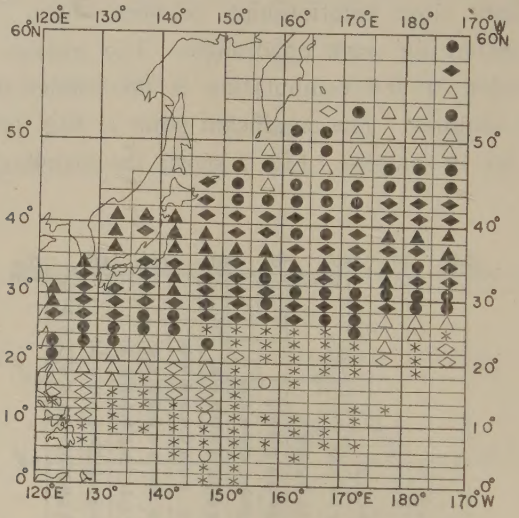

Fig. 3. Probable errors of the linear relation between the change of monthly sea surface temperature and the heat exchange between the sea and the atmosphere.

▲≧1.0°C	0.4>◇≧0.2°C	
1.0>◆≧0.8	0.2>○	
0.8>●≧0.6	*	insignificant correlation at 95% level
0.6>△≧0.4	blank	not computed

of the linear relation obtained as the regression equation. The error is 0.6—1.0°C in the region between 26°N and 48°N in general and is largest in many quadrangles between 30°N and 40°N. The quadrangles with large errors show marked fluctuations in ΔT, but the ratio of the error to the standard deviation of ΔT is not so large as compared with that for quadrangles with small errors.

4. Discussion

The change of the water temperature of the surface layer is given by

$$\rho c_p \frac{\partial \theta}{\partial t} = -\rho c_p \vec{V} \cdot \nabla\theta + \rho c_p K_H \dot{\nabla}_H{}^2 \theta + \rho c_p K_V \frac{\partial^2 \theta}{\partial z^2} - \frac{\partial Q'}{\partial z}, \qquad (1)$$

where θ is water temperature, $\vec{V}$ current velocity, Q' incoming heat energy, ρ

density, c_p specific heat at constant pressure, K_H horizontal diffusion coefficient, K_V vertical diffusion coefficient. z is the vertical coordinate directed downward and t the time coordinate. If Q' vanishes at the depth D, the integration of (1) from $z=0$ to $z=D$ leads to

$$\rho c_p \frac{\partial \Theta}{\partial t} = -\rho c_p \int_0^D \vec{V}_H \cdot \nabla_H \theta dz + \rho c_p K_H \int_0^D \nabla_H^2 \theta dz - \rho c_p \int_0^D w \frac{\partial \theta}{\partial z} dz + \rho c_p K_V \frac{\partial \theta}{\partial z}\Big|_{z=D} + Q, \tag{2}$$

with the boundary conditions at the sea surface ($z=0$),

$$\rho c_p K_V \frac{\partial \theta}{\partial z} = Q_b + Q_h + Q_e$$

$$Q' = Q_s.$$

Here,

$$Q = Q_s - (Q_b + Q_h + Q_e),$$

$$\Theta = \int_0^D \theta dz.$$

Q_b, Q_h and Q_e are defined in Section 2 and $\vec{V}_H$ is the horizontal component vector of velocity and w is the vertical component of velocity. The first and the second terms in the righthand side of (2) means the horizontal heat exchange in the layer and the third and the fourth terms the vertical heat exchange with the lower layer and the last term the heat exchange between the sea and the atmosphere. Though the horizontal heat exchange fluctuates, the seasonal variation is removed and the fluctuation is not so marked as that of Q in general.

Now, it is assumed that $\partial\theta/\partial t$ is independent of depth. Though this is not reasonable, it is accepted to the study on the relation of the first approximation. In addition, the fluctuation of the vertical heat exchange is assumed to be not so marked. Then, high correlation is expected between the change of sea surface temperature and Q. Then, the following relation is obtained.

$$\frac{\partial \theta}{\partial t} = \alpha Q + \beta$$

Here,

$$\alpha = 1/\rho c_p D$$

$$\beta = Q_i/\rho c_p D$$

$$Q_i = -\rho c_p \int_0^D \vec{V}_H \cdot \nabla_H \theta dz + \rho c_p K_H \int_0^D \nabla_H^2 \theta dz - \rho c_p \int_0^D w \frac{\partial \theta}{\partial z} dz + \rho c_p K_V \frac{\partial \theta}{\partial z}\Big|_{z=D}.$$

Q_i is called the internal heat exchange in this report and its positive value

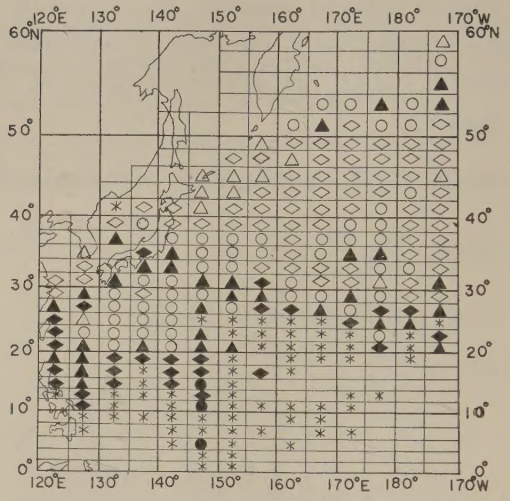

Fig. 4. Depths of the surface layer D.

●	≧200m	60>◇≧40m	
200>◆≧100		40>△	
100>▲≧ 80		*	insignificant correlation at 95% level
80>○≧ 60		blank	not computed

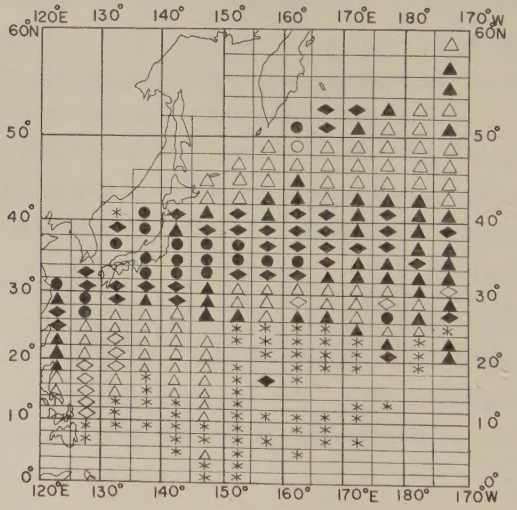

Fig. 5. Internal heat exchange Q_i in the surface layer.

○≦−0.10 ly/min		0.05<◆≦0.10 ly/min	
−0.10<◇≦−0.05		0.10<●	
−0.05<△≦ 0.00		*	insignificant correlation at 95% level
0.0 <▲≦ 0.05		blank	not computed

means heat gain. α and β are obtained from the regression equation between ΔT and Q and then, the depth D and the internal heat exchange Q_i are evaluated. Such assumptions may be applied to the quadrangles of significant correlation in Fig. 2. Those values are shown in Figs. 4 and 5.

The depth D is smaller than 100 m in general to the north of 20°N. Between 40°N and 50°N, the depth is 40—50 m in most quadrangles. This depth is the depth of the layer affected by seasonal variation of the heat exchange between the sea and the atmosphere, and is subject to marked seasonal variations, but the values shown here may be interpreted as annual means of the depth. Then, the feature of the distribution is compatible with the observed results. The values are larger than those computed by WYRTKI. His values are in good agreement with the depth of the mixed layer in summer when the mixed layer has the minimum thickness. Hence, it is natural that the values shown here are larger than his results. The value is too large in the southern part and, in some quadrangles, the depth is much different from the ones in the neighbouring areas. It is not wise to apply these assumptions to those quadrangles.

The internal heat exchange is positive in the region where the Kuroshio flows, that is, heat energy is supplied from the surrounding area by advection and horizontal diffusion and, vertically, the energy is given by sinking, because $\partial\theta/\partial z$ is generally negative in the layer. Though the contribution of the vertical motion can not be estimated at present, the supply of the heat energy is supposed to be mainly due to advection and horizontal diffusion, for the heat gain is large in the region where the horizontal gradient of surface temperature is large and the current speed is high. The heat energy supplied by horizontal exchange is given to the atmosphere. The extreme value of the internal heat exchange is more than 0.10 ly/min and is found in the region where much heat energy is supplied to the atmosphere from the sea. On the other hand, the heat energy is carried away in horizontal directions where the sea gains the heat energy by air-sea interaction. Such situations are found south of 30°N and north of 40°N. And, in the Kuroshio region, heat loss by internal exchange is found in the sea east of Formosa. Since current speeds are high in the area, most of the heat loss is due to advection. The correspondence of internal heat exchange to the heat exchange between the sea and the atmosphere results from Equation (2). Though the value is considered to have no small error, the outline of the distribution is reasonable.

It is interesting that significant correlations are not found in the southern

part, where the fluctuations of the change of monthly sea surface temperature and of the heat exchange between the sea and the atmosphere are not marked as mentioned in the previous section. The simple model employed above can not be applied to such regions. Factors other than Q are important and various models are employed. If $\partial\theta/\partial t$ is independent of depth as before and the fluctuation of w is important, the change of sea surface temperature will be highly correlated with w from Equation (2). For example, SUDO (1960) found the existence of strong divergence at the sea surface in the southern part, and deduced upwellings in the region on the basis of continuity and the distribution of the surface currents. In Fig. 2, correlation coefficients are not significant in the area 150—170°E and south of 20°N, an area of strong divergence and upwelling according to SUDO's study. In such a case, the upwelling is more effective than the heat exchange between the sea and the atmosphere, and the correlation with the vertical motion will be significant.

From the description noted above, it is referred that the simple computation represents reasonable distribution of the depth of the surface layer and of the internal heat exchange as the first approximation for the area of significant correlation. This is one of the evidences that the change of the sea surface temperature is dependent on the fluctuation of the heat exchange between the sea and the atmosphere. For the area of insignificant correlation, on the other hand, other models are to be employed to obtain high correlation.

5. Concluding remarks

The change of monthly sea surface temperature in the northern part of the Western North Pacific is highly correlated with the heat exchange between the sea and the atmosphere and is represented as a linear function of the heat exchange. The probable error is at most 1°C. This is accepted as the first approximation. The depth of the surface layer and the internal heat exchange are also obtained as accompanied with the procedure. If they are regarded as their annual means, those distributions are reasonable and support the relation from oceanographic considerations.

The assumptions are not appropriate to more detailed discussions and the method should be improved as the following step. Besides, it is to be examined which factor plays the predominant role in the change of monthly sea surface temperature and which factors are closely related with the deviation from the linear relation derived in this report.

Acknowledgement: The author wishes to express his thanks to Dr. T. NAN'NITI for his valuable advices and to Mr. K. NISHIYAMA and Miss K. OKADA of Meteorological Research Institute for their help in the study.

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Some problems relating to fluctuation of hydrographic conditions in the sea northeast of Japan (Part 2)

—Fluctuation of the warm eddy cut off northward from the Kuroshio—

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(Received June 13, 1969)

Abstract: The fluctuation of water mass distribution, especially the formation of the warm eddy from the Kuroshio, is discussed from many distributions of oceanographic elements. In this study are mainly used the data of hydrographic observations by the Hakodate Marine Observatory, the Japan Meteorological Agency, the Tohoku Regional Fisheries Research Laboratory and the Japan Maritime Safety Agency from 1954 to 1955 and from 1959 to 1966.

The results are as follows:

1. The fluctuation of the warm eddy depends upon many factors—temperature at the depth of 100 m, the volume transport, the velocity of surface current, and the diameter; the relative importance of these factors is discussed.

It is analyzed that the life time of the warm eddy with the volume transport which amounts to $20 \times 10^6 m^3$/sec or more is more than than one year. Occasionally the values of the large warm eddies (Feb.-Mar., 1954, Aug.-Sept., 1960) are more than $30 \times 10^6 m^3$/sec, which corresponds to about 60% of eastward flow of the Kuroshio, and the velocity of surface current around them is about 3 knots and the diameter is about 150 nautical miles. These warm eddies are stable and durable. But the warm eddies with the volume transport less than $10 \times 10^6 m^3$/sec, velocity about 1 knot and the diameter less than 80 nautical miles, are apt to move northeastward, and to become weaker or diffuse rapidly.

2. It may be concluded that the increase of amplitude in meander of the Kuroshio and the formation of eddy cut off from the Kuroshio southeast of Kinkasan are caused mainly by meteorological disturbances.

1. Introduction

The Kuroshio east of Honshu has been studied by many hydrographers of Japan (UDA, SAITO, ICHIYE, SUGIURA, KOIZUMI, NAN'NITI, MASUZAWA, HANZAWA, FUKUOKA, KIMURA and KAWAI). They suggest that there are some remarkable seasonal and yearly fluctuations in the location of the boundary between the

Kuroshio and the Oyashio.

Especially, MASUZAWA (1955) analyzed the general features of the current-pattern of the Kuroshio and its branching and the warm eddies in the neighbourhood of the stream axis by the observations from 1933 and 1942 and from 1954 to 1955.

The movement of the small cold eddy along the Inner Cold Zone in the left side of the primary Kuroshio Front was discussed from late spring to early summer in 1959 by KAWAI (1961). And it was pointed out that the propagation velocity of characteristic pattern of the Kuroshio Front was about 0.5 knot. The multiple ship survey in the vicinity of the Kuroshio Front in 1957 shows that the Kuroshio Front moved northward as a rate of about 12 nautical miles a day, namely the speed of 0.5 knot (MASUZAWA, 1958)

The cut-off of the warm eddy from the Kuroshio in this region has been observed many times. In the recent years, KIMURA (1962, 1968) studied the movement of the warm eddies and pointed out that the cut-off occurs frequently in May and November.

The volume transport of the warm eddy east of Kinkasan (lat. 38°15′N) was calculated by several authors (SUGIURA, 1954, 1955; MASUZAWA, 1955) and the values between 10 and $30\times10^6 m^3$/sec were obtained.

It was pointed out theoretically by ICHIYE (1955) and UDA (1968) that by the disturbing forces of travelling cyclones the Kuroshio became unstable and its amplitudes of meandering increased.

In this paper some close relations between the warm eddy and the Kuroshio and between the fluctuations of hydrographic conditions and meteorological disturbances are explained.

2. On the volume transport

Figure 1 shows the schematic charts of the volume transport of different currents. The lines with arrows indicate the approximate direction of the transport above 1,000 m for the Kuroshio and the warm eddy, above 600 m for the Oyashio and above 400 m for the Tsushima Current. The arrows with dash lines show cold currents, and those with solid lines show warm currents. The volume transports in the sea northeast of Japan were computed by several authors, between 40 and $60\times10^6 m^3$/sec for the main flow of the Kuroshio, and between 10 and $30\times10^6 m^3$/sec for the branch of the Kuroshio. A remarkable warm eddy east of Kinkasan is seen (SUGIURA, 1955; MASUZAWA, 1955). The values between

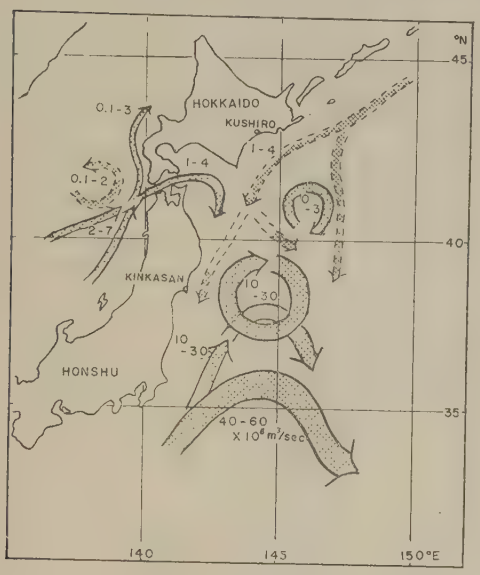

Fig. 1. Schematic chart of volume transport in the sea northeast of Japan. The lines with arrows indicate the approximate direction of the transport above 1000 m for the Kuroshio and the warm eddy, above 600 m for the Oyashio and above 400 m for the Tsushima Current.

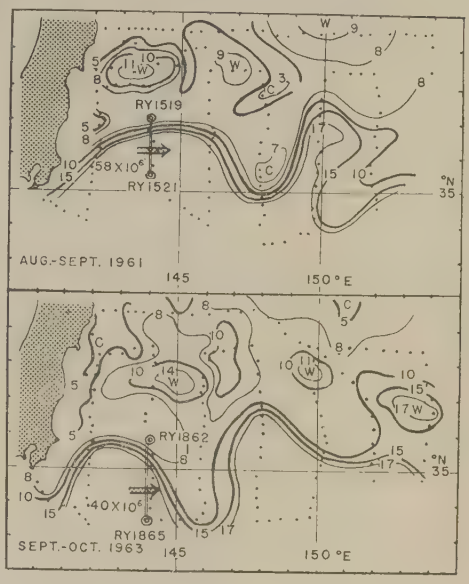

Fig. 2. Distributions of temperature at the depth of 200 m and volume transport of main flow of the Kuroshio referred to the 1000 d. b. surface in Aug.-Sept., 1961 and Sept.-Oct., 1963 by R/V *Ryofu Maru*.

2 and 7×10^6m^3/sec for the Tsushima Current and between 1 and 4×10^6m^3/sec referred to the 400 d.b. surface for the Tsugaru Current were obtained by HATA (1962). The values between 1 and 4×10^6m^3/sec referred to the 600 d.b. surface for the coastal branch of the Oyashio southeast of Hokkaido and between 0.2 and 3×10^6m^3/sec for the small warm eddy southeast of Kushiro (lat. 42°59'N) were obtained by SUGIURA (1954, 1955) and HATA (1965, 1968).

There are large fluctuation in the volume transport of the Kuroshio and others. For example, as seen in Figure 2, the volume transport of main flow of the Kuroshio between Ry 1519 (36°59'N) and Ry 1521 (35°11'N) along 144°E is 58×10^6m^3/sec referred to the 1,000 d.b. surface in August—September, 1961, and in September—October, 1963 the value between Ry 1862 (35°46'N) and Ry 1865 (33°28'N) is 40×10^6m^3/sec, which is 69% of that in 1961.

3. On the warm eddy east of Kinkasan

It was pointed out by MASUZAWA (1955) that the warm eddy with the size

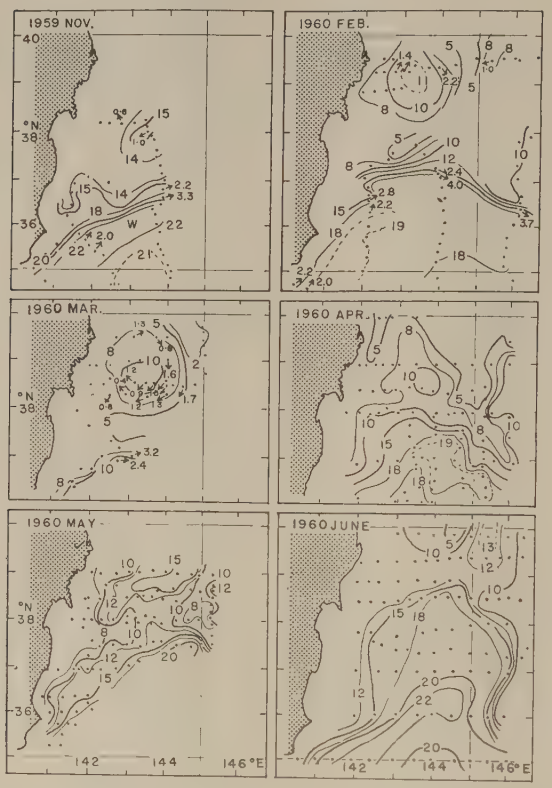

Fig. 3-1. Distributions of temperature at the depth of 100 m and surface currents with GEK from Nov., 1959 to June, 1960.

of 100 miles or more in diameter seemed to be unable to remain completely isolated in the cold region in the period from November, 1954 to March, 1955.

Figures 3 and 4 show the distributions of temperature at the depth of 100 m and of surface currents with GEK from November, 1959 to May—June, 1961, and from September to December, 1966, respectively, though the distribution of temperature in December, 1960 is the one at the sea surface (Figure 3-2).

As seen in Figure 3, the main Kuroshio Current flowed eastward between 36°N and 37°N in the period from November, 1959 to May, 1960 and then increased their amplitude of meandering in June, 1960 and the Kuroshio Current showed a remarkable loop-shaped path in July, 1960 similar to the distribution in August, 1942 (Masuzawa, 1955).

Another example from September to November, 1966 are shown in Figure

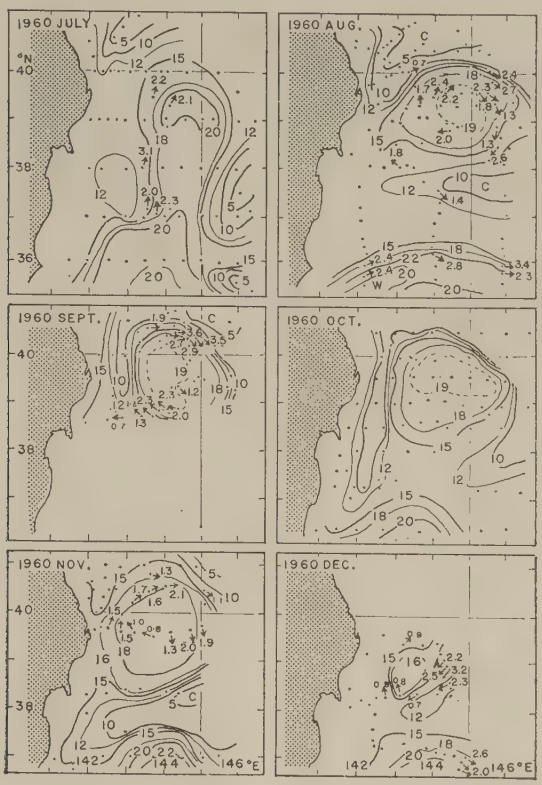

Fig. 3-2. Distributions of temperature at the depth of 100 m and surface currents with GEK from July to Nov., 1960. The distribution of temperature in Dec., 1960 is the one at the sea surface.

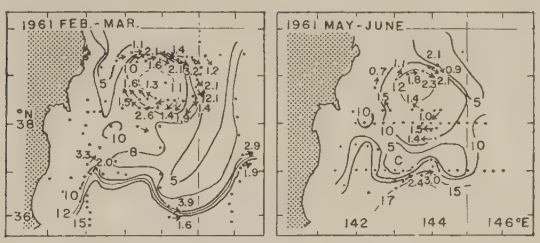

Fig. 3-3. Distributions of temperature at the depth of 100 m and surface currents with GEK in Feb.-Mar. and May-June, 1961.

4. An isolated anticyclonic large eddy cut off from the main Kuroshio Current is formed in this region.

Figure 5 shows the distribution of temperature at the depth of 200 m and

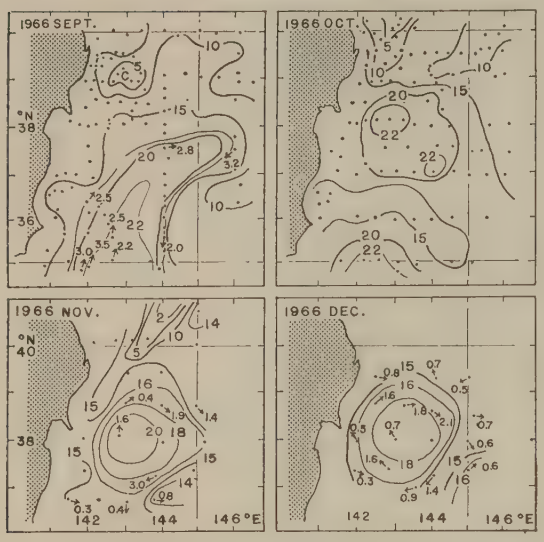

Fig. 4. Distributions of temperature at the depth of 100 m and surface currents with GEK from Sept. to Dec., 1966.

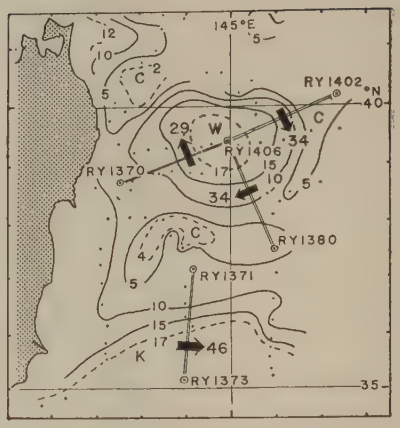

Fig. 5. Distribution of temperature at the depth of 200 m and volume transports ($\times 10^6 m^3$/sec) referred to the 1,000 d.b. surface in Aug. 13-Sept. 4, 1960. K: Kuroshio, C: Oyashio, W: warm eddy.

the volume transport referred to the 1,000 d.b. surface from August 13 to September 4, 1960. The volume transport of main flow of the Kuroshio is $46\times 10^6 m^3$/sec, and that of the warm eddy amounts to $29—34\times 10^6 m^3$/sec, which corresponds to about 70% of the Kuroshio Current. And the velocity of surface currents with GEK of the warm eddy is between 2.0 and 3.6 knots, its diameter shown by 10°C-isotherm at the depth of 200 m is about 150 miles, and the temperature and salinity at the depth of 100 m are 19°C and 34.83‰ in the center of the eddy. Such a remarkable large warm eddy as this was not found east of Kinkasan from 1955 to 1966.

Masuzawa (1955) analyzed that the remarkable isolated warm eddy in March, 1954 with the volume transport which amounted to about $30\times 10^6 m^3$/sec connected the main Kuroshio Current in May, and then in summer the connection became estranged. The warm water mass became a completely isolated warm eddy in autumn and furthermore it was almost vanishing by diffusion in March, 1955.

From August, 1960 to May, 1961 the warm eddy is isolated almost from the main Kuroshio Current as shown in Figures 3-1, 3-2 and 3-3.

As compared the structure of the warm eddy in August—September, 1960 with that in February—March, 1961, the volume transport of the former amounts to about $29—34\times 10^6 m^3$/sec and the value of the latter five months later de-

creases to about 23—26×10⁶m³/sec which is about 75% of the former. The decrease of volume transport is less than expected. There are no remarkable changes in the surface currents of the warm eddy, 2.0—3.2 knots, in the period. The temperature at the depth of 100 m decreases from 19°C to 11°C, but the temperature at the depth of 400 m is about 10°C in both cases (refer to Figure 8), and the diameter decreases to a half of that in August—September, 1960.

In October, November and December, 1960, the hydrographic observations were restricted to shallow depths, so that the value of the volume transport of warm eddy referred to the 600 d.b. surface was only 18×10⁶m³/sec, but the values referred to 1,000 d.b. surface might be larger than about 23×10⁶m³/sec.

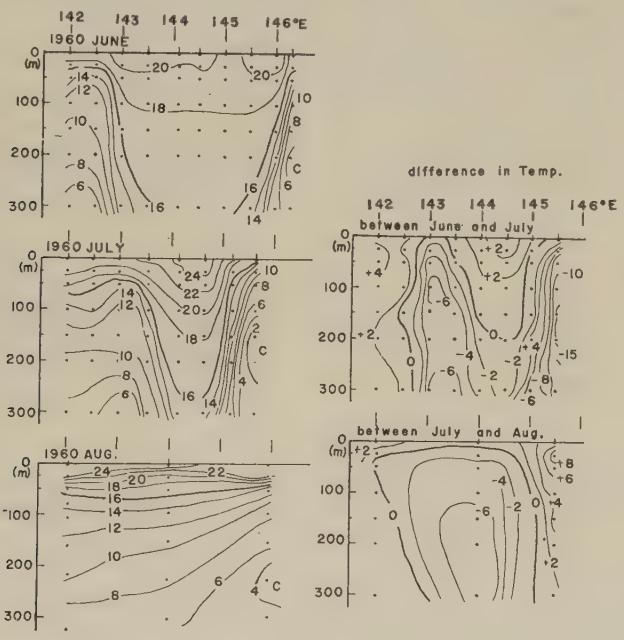

Fig. 6. Vertical distributions of temperature from June to Aug., 1960 and the distributions of the difference between June and July, between July and Aug. along 38°N.

Figure 6 (left) shows the vertical distribution of temperature from June to August, 1960 along 38°N. In June the width of the Kuroshio water is about 150 miles and in July the pure Oyashio water colder than 2°C is seen at 145.6°E at depths of 150 m and 200 m, and the width of Kuroshio water decreases to about 80 miles, which is remarkably narrower than in June. In August the main

Kuroshio water do not appear in the section (refer to Figure 3-2), and the large change of the hydrographic conditions is seen in the period.

In order to see the variation of the Kuroshio water mass, the vertical distributions of the difference in temperature between June and July, and between July and August along 38°N are illustrated in Figure 6 (right). Between June and July, the area of increase in temperature exists near 142°E and 144.5°E and the values are 4°C in the surface layer, but the other parts show decrease. Between July and August, except in the near surface layer and in the neighbourhood of 145.5°E, the temperature shows decrease (-2—6°C).

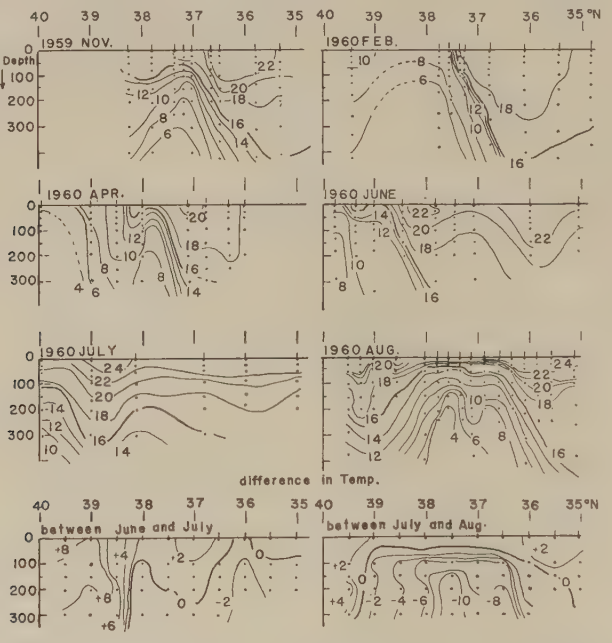

Fig. 7. Vertical distributions of temperature from Nov., 1959 to Aug., 1960 and the distributions of the difference between June and July, between July and Aug., 1960 along 144°E.

Figure 7 shows the vertical distributions of temperature from November, 1959 to August, 1960 along 144°E. In November, 1959 the position of the Kuroshio Front exists at about 36.5°N, in February, 1960 about 37°N, in April about 37.5°N, in June about 38.5°N and then in July about 40°N. Thus, the Kuroshio Front moves gradually northwards, and there seems to be a very considerable change of the position from June to July. The Kuroshio Front moves northward at a rate of about 3 nautical miles a day in the period from June to July.

As seen in the current pattern (Figure 3) and in the vertical distribution along 144°E in August (Figure 7), the center of warm eddy exists at about 39°N and the Kuroshio Front lies father south, 36°N, in August than in July. That is, the Oyashio water is not seen in the region south of 38°N in June and July, but the penetration of the Oyashio water in which the temperature is 4—8°C is seen about 37.5°N at depths of 150 m and 400 m in August, 1960.

The vertical distributions of difference in temperature between June and July, and between July and August along 144°E are illustrated at the bottom of Figure 7. Between June and July, the area of increase in temperature exists between 38°N and 40°N and the increase amounts to 4—8°C, and the area of decrease exists at 36°N. The area of increase is larger than that of decrease. On the contrary, between July and August the area of decrease is larger than that of increase, and the area of decrease exists from 36°N to 39°N at depths of 100 m and 300 m and the decrease amounts to 2—10°C.

From these facts, it is verified that the speedy change in the pattern of the Kuroshio water occurs from July to August, 1960 in this region, as will be mentioned in § 4, these changes may be associated with meteorological disturbances.

The lower right of the Figure 8 shows location of the serial stations from June, 1960 to May, 1961. The hydrographic data at the representative stations at the center of warm eddy and in the Kuroshio area in the upper 600 m layer are shown in Table 1. The upper of the Figure 8 shows the profiles of temperature and salinity in Table 1. From June to July, 1960, the temperature in the crest of the meandering path of the Kuroshio seems to have

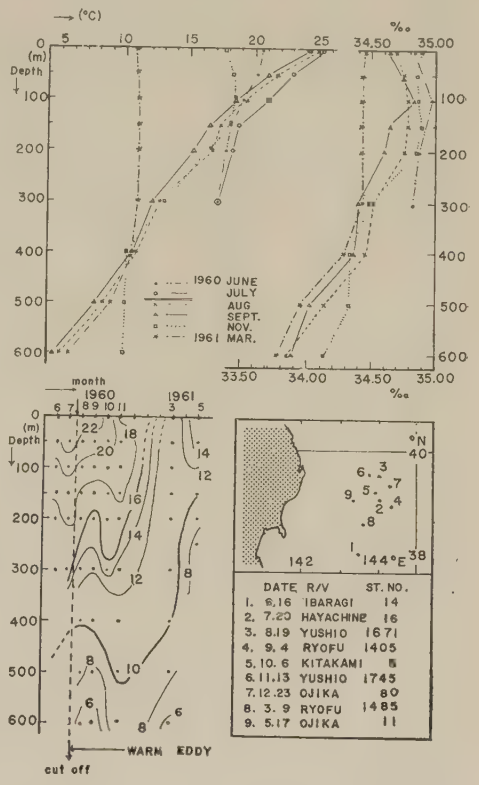

Fig. 8. Profiles of temperature and salinity shown in Table 1 (upper), isopleth of the temperature in the warm eddy from June, 1960 to May, 1961 and serial station locations (lower).

Table 1. Hydrographical data in the upper 600 m layer at the center of the warm eddy and the Kuroshio water.

Warm eddy

Date	1960. 8.19		1960. 9.4		1960. 11.13		1961. 3.9.	
St.	Yu1671		Ry1405		Yu1745		Ry1485	
	39°30′N	143°59′E	38°50′N	144°21′E	39°30′N	143°45′E	38°30′N	143°33′E
Depth	(°C)	(‰)	(°C)	(‰)	(°C)	(‰)	(°C)	(‰)
0	25.0	34.69	24.3	34.47	17.7	34.89	10.8	34.45
100	18.29	.79	18.46	.83	18.5	.85	10.93	.43
200	16.62	.57	15.26	.60	16.70	.83	10.97	.43
300	12.60	.51	12.03	.40	12.96	.49	10.88	.42
400	10.93	.45	10.69	.36	10.08	.34	10.14	.29
500	8.25	.13	7.04	.02	10.04	.33	8.08	33.95
600	4.82	33.84	4.55	33.87	9.06	.31	5.82	.77

Kuroshio water

Date	1960. 2.19		1960. 6.16		1960. 8.16	
St.	Ry1311		14		Ry1374	
	36°41′N	144°05′E	37°57′N	144°30′E	35°08′N	143°58′E
Depth	(°C)	(‰)	(°C)	(‰)	(°C)	(‰)
0	18.2	34.96	21.0	34.84	26.6	34.96
100	18.35	.94	19.04	.99	19.21	.81
200	17.51	.76	17.77	.85	17.77	.87
300	16.17	.74	16.98	.83	16.24	.70
400	13.94	.54			13.94	.54
500	11.21	.42			13.81	.51
600	7.42	.15			10.96	.24

increased in the upper 200 m layer, but at the depth of 300 m there are no changes of the temperature which is remained in about 17°C. The highest of the salinity, 34.99‰, appears at the depth of 100 m in June.

The distributions of temperature and salinity at the center of the warm eddy seem to be unchanged from August to November, 1960 except in the layers shallower than 50 m and deeper than 500 m. From September to November the temperature decreases and salinity increases, and these changes of temperature and salinity in the surface layer (0 m—50 m) may be explained mainly by cooling of the surface water due to the evaporation. In the layer deeper than 500 m the temperature and salinity increase in the period, and it may be due to the sinking of warm water by the anticyclonic circulation.

Table 2. Variations of characteristic properties of warm eddies and the Kuroshio water.

Year	Month	Warm eddies off Kinkasan: Scale	Max. Temp. 100m (°C)	Diameter (mile)	Surface Current north (knot)	Surface Current south	Volume Transport ×10⁶ (m³/sec)	Main stream of the Kuroshio along 144°E: Position (°N)	Surface Current (knot)	Volume Transport ×10⁶ (m³/sec)	Position of Stations (°N)	Remark
1953	11–12		20					38.2				
1954	2– 3	La.	16	150–200			33 (28–35)	36.0		53	33.5–37.1	( ) Masuzawa (1955)
〃	5		18	–150–			(27–29)	36.5		43	35.1–38.0	
〃	8		18	〃			(22–25)	36.5		45	35.5–37.0	
〃	11–12	Mid.	16	150–180			16 (15–21)	37.0		55	35.0–37.3	
1955	2– 3	Sm.	6	–60–			5	37.5		48	35.0–38.3	
1959	11	Mid.	15	60–100	2.2	—		36.5	3.3	51	34.7–37.0	
1960	2	〃	11	60– 80	2.2	—	16	37.2	4.0	50	36.1–37.8	
〃	3	〃	10	60–100	2.2	1.4		37.2	3.2*			* along 143°E
〃	4	〃	10	–80–				37.5				
〃	5	Sm.	15	〃				37.5**				** along 143.5°E
〃	6	〃	13	〃				38.8				
〃	7	〃	12	〃				39.5	3.1			
〃	8	La.	20	130–180	2.7	2.0		36.3	2.8	46–50	34.0–37.1	
〃	9	〃	19	〃	3.5	2.3	29–34					
〃	10	〃	19	–150–				36.5				
〃	11	〃	18	〃	2.1	2.0		37.0				
〃	12	Mid.	16***	–100–		3.2						*** surface temp.
1961	2– 3	〃	11	〃	3.2	2.6	23–26	36.5	4.5	50	35.0–37.0	
〃	5– 6	Sm.	12	–80–	2.3	1.4		36.8				
〃	8	〃	11	〃	1.6	1.2		35.8	3.0	58	35.2–37.0	
1962	2– 3	Mid.	15	–100–	2.2	1.0	12	37.3	2.3	48	34.7–37.0	
〃	8– 9	〃	15	〃	1.4	1.7	17	37.2	4.3	52	35.6–37.5	
1963	2– 3	Sm.	8	40– 60				36.8	3.8	53	35.2–37.7	
〃	9	Mid.	14	80–100	2.7	1.8	14	35.5	3.6	40	33.5–38.0	
1964	1– 2	Sm.	10	50–100	0.5	1.5				45	33.5–38.0	
1965	10	La.	22	–150–								
1966	2– 3	Mid.	9	50–100			16	36.5	3.4	48	35.3–38.0	
〃	11–12	〃	20	100–150	1.9	3.0	24					

La. (large) Mid. (middle) Sm. (small)

In March, 1961, the profiles between 300 m and 400 m almost coincide with the former (from August to November, 1960), but the changes in the layer shallower than 200 m and in the layer deeper than 500 m may be caused by the diffusion from the cold water of the Oyashio and cooling.

In the warm eddy, the smallest change of temperature and salinity are observed in the layers of 300 m and 400 m from August, 1960 to March, 1961.

The lower left of the Figure 8 shows the isopleth of the temperature in the layer shallower than 600 m in the area of the Kuroshio water from June to July and near the center of the warm eddy from August, 1960 to May, 1961. The depth of 16°C-isotherm is deeper than 300 m in the crest of the meandering path of the Kuroshio water in June and July, but in the warm eddy the depth becomes shallower from August to September immediately after cut-off from the main stream and then the depth becomes again deeper in October. This tendency is also seen in the depth of 10°C-isotherm, which is deepest in November, 1960, approximately 500 m, and as above explained, it is formed by sinking due to anticyclonic circular current. In March, 1961, the 10°C-isotherm exists at the depth of 400 m, and furthermore in May, 1961 it becomes shallow rapidly to the depth of 150 m. It shows the decreasing of the thickness of the warm eddy by diffusion. The large warm eddy in August, 1960 becomes the small warm eddy in May—June, 1961.

Figure 9 shows the T-S curves of the six stations near the center of the warm eddy and in the Kuroshio water from February, 1960 to March 1961. In

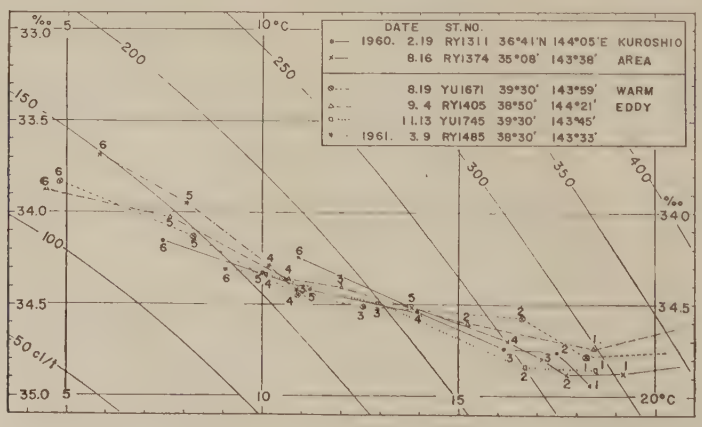

Fig. 9. Relations between temperature and salinity at the warm eddy and the Kuroshio water in Feb., 1960—Mar., 1961.

order to see relations between the warm eddy and the Kuroshio water mass and the change of the eddy in the period are shown in Figure 9. These stations were taken by the R/V *Yushio Maru* and *Ryofu Maru*. At the stations Ry 1311 and Ry 1374, the Kuroshio water mass at the depths between 100 m and 600 m is characterized by a nearly straight T-S curve between the point T=9°C, S=34.2‰ and T=16°C, S=34.9‰. At the stations Yu 1671, Ry 1405 and Yu 1745 the Kuroshio water in the warm eddy is seen between T=6°C, S=33.9‰ and T=18°C, S=34.8‰. The water in the eddy is a little lower in temperature and salinity than the Kuroshio water, but the curves in eddy almost agree with those in the Kuroshio except the surface layer and the layer deeper than 500 m, in spite of such a long duration from August, 1960 to March, 1961. The warm eddy is connected with the main Kuroshio water only in October, in the other period it is almost isolated. In March, 1961 the water mass at the station Ry 1485 (refer to Table 1) shows the homogeneous water-mass in the layer shallower than 300 m, and the temperature and salinity are about 11°C and 34.45‰, and by cooling with evaporation and diffusion these values decrease by about 7°C and 0.4‰ from those in November, 1960.

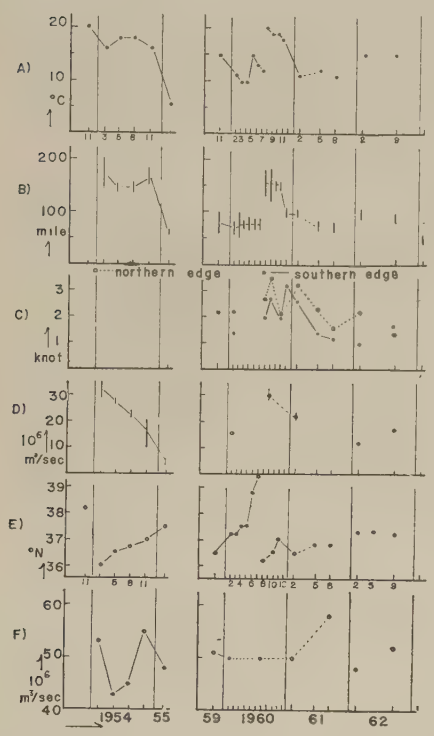

Fig. 10. Variations of characteristic properties of warm eddies and the Kuroshio water. A) Maximum temperature of the warm eddy at the depth of 100 m. B) Diameter of the warm eddy. C) Surface currents in the northern and southern edge of the eddy. D) Volume transport of the eddy. E) Position of northern edge of the Kuroshio water along 144°E. F) Volume transport of main flow of the Kuroshio.

Figure 10 shows the variation of characteristic properties of the warm eddies and the Kuroshio water, that is, the maximum temperature at the depth of 100 m in the warm eddy, the diameter of the warm eddy, the velocities of surface current in northern and southern edge of the eddy, the volume transport of the warm eddy, the position of northern edge of the Kuroshio Front along 144°E, and the volume transport of main flow of the Kuroshio in 1953—1955 and

1959—1962.

As for the relation between the two large warm eddies in 1954 and in 1960, the former is almost vanishing by diffusion after about four months and the temperature at the depth of 100 m is about 6°C, the diameter is about 60 miles, the volume transport amounts to about $5\times10^6 m^3$/sec in March, 1955. But the latter continues March, 1961 and the temperature at the depth of 100 m is about 11°C, the diameter is about 100 miles, the velocity is about 3 knots and the volume transport amounts to about $23\times10^6 m^3$/sec.

The cold water of the Oyashio lies farther south from 1954 to 1955, but from 1960 to 1961 lies farther north than normal (HATA, 1969). It is verified that the warm eddy do not diffuse very much with the cold water of the Oyashio from November, 1960 to March, 1961, but the warm eddy from November, 1954 to March. 1955 becomes weak or diffuses rapidly (MASUZAWA, 1955).

When the velocity of the surface current of the warm eddy in this region is about 3 knots, the eddy is not so weak, the eddy is maintained more than one year unless the cold water of the Oyashio penetrates into this region.

In this region the data of the areal observations and the repeated observations in very short intervals have not been so much accumulated that immediate relation between the fluctuation of volume transport of the Kuroshio and the warm eddy is obscure.

4. Relation between the fluctuation of hydrographic condition and the meteorological disturbance

On the relation between the fluctuation of hydrographic condition and meteorological disturbance, it is pointed out theoretically by ICHIYE (1955) that the increase of amplitude in meandering of the Kuroshio south of Honshu is influenced by the disturbing forces of the travelling cyclones.

Figures 11 and 12 show the tracks of cyclones in the neighbourhood of Japan and the schematic charts of the hydrographical condition in the neighbourhood of Polar Frontal Zone from May to August, 1960 and 1961. As seen in Figure 11, there is a difference between 1960 and 1961, especially in May and August. That is, four cyclones travel in this region off Kinkasan in May, 1960, while the passage travelling cyclones in this region is not seen but three remarkable cyclones travel on the Japan Sea in May, 1961.

In Figure 12 it seems to increase the amplitude of meandering of the Kuroshio southeast of Kinkasan from May to July, 1960, therefore, the increase of

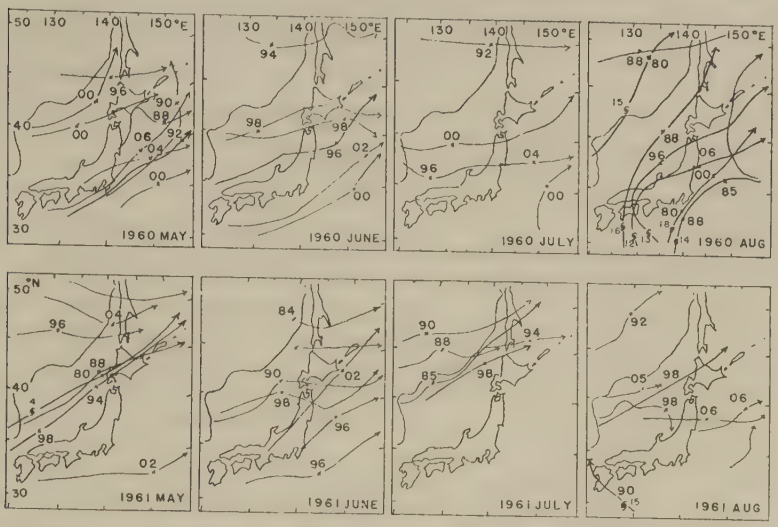

Fig. 11. Tracks of cyclones from May to Aug., 1960 and 1961.

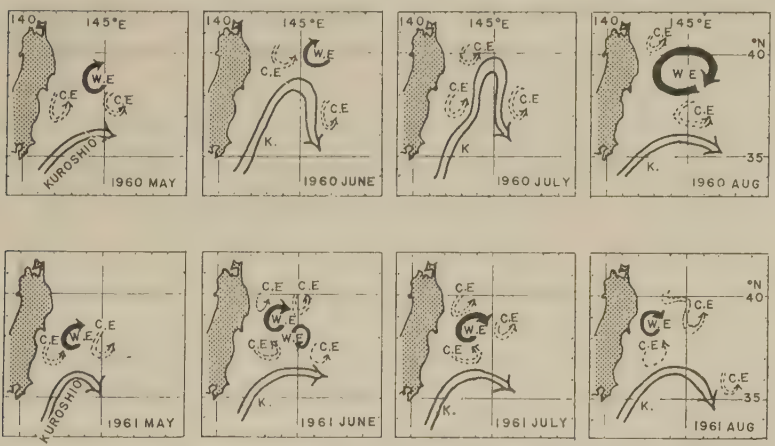

Fig. 12. Schematic representation of the hydrographic condition in the neighbourhood of the Polar Frontal Zone from May to Aug., 1960 and 1961.
W.E: warm eddy C.E: cold eddy

amplitude in meandering of the Kuroshio may be influenced by these travelling cyclones, but this phenomenon is not seen in 1961. In August, 1960 three typhoons (No. 12, No. 14 and No. 18) travel in this region and Typhoon No. 14 is most remarkable and the lowest atmospheric pressure shows about 985 mb., and the large warm eddy may be formed by the cut-off from the ridge nearest the land (143°E –146°E) in meandering of the Kuroshio by the meteorological disturbance.

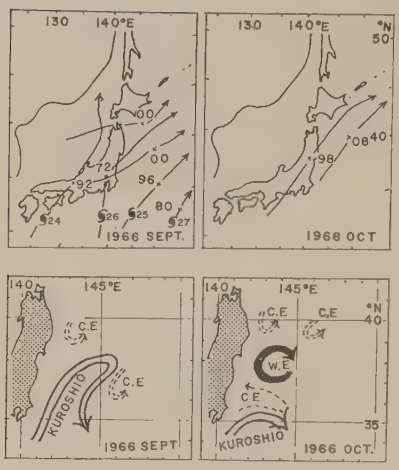

Fig. 13. Tracks of cyclones (upper), schematic representation of the hydrographic condition from Sept. to Aug., 1966 (lower).
W.E: warm eddy C.E: cold eddy

In 1961 many cyclones or typhoons do not pass over the region concerned and a large warm eddy is not formed in this region.

The upper of Figure 13 shows the tracks of cyclones and the lower shows the schematic chart of the hydrographical condition from September to October, 1966. It is pointed out by UDA (1968) that the cut-off from meandering of the Kuroshio in the period from September to October, 1966 was caused in the region by the meteorological disturbance.

It may be concluded that the increase of amplitude in meandering of the Kuroshio and the formation of the large warm eddy is associated with the meteorological disturbances in the sea southeast of Kinkasan.

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Distribution of AOU in the surface waters of Kuroshio and Oyashio and its oceanographical significance —The delay of gas exchange between air and sea

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Abstract: A consistent trend recognized as to the distribution of AOU in the surface water of the Kuroshio or the Oyashio has been illustrated as an evidence to show that the rate of gas exchange is slower than that of heat exchange between air and sea.

1. Introduction

AOU, which is the abbreviation of apparent oxygen utilization, is defined as the difference between the saturation amount at *in situ* water temperature and chlorinity under one atmosphere of air and the observed amount of oxygen. The value of the saturation amount of oxygen in sea water has recently become controversial. One employs the amount of oxygen saturated under the dry air and another, that under the wet air. Furthermore, there are found some discrepancies in the solubility measurements among several workers (Carritt and Carpenter, 1966). In the Japan Meteorological Agency, the Fox's solubility values have long been in use. They are the saturation amounts under the dry air of one atmosphere.

In this paper, the conspicuous features in the distribution of AOU based on the Fox's saturation amount has been pointed out and its oceanographical significance discussed.

2. Distribution of AOU in the surface waters along the axis of the Kuroshio

Fig. 1 shows two groups of sampling sites, one closer to and another farther from the source region of the current. Fig. 2 shows the temperature vs. chlorinity diagram pertaining to both waters collected at two sites mentioned above. The chlorinity range covered by the individual waters closer to the source region

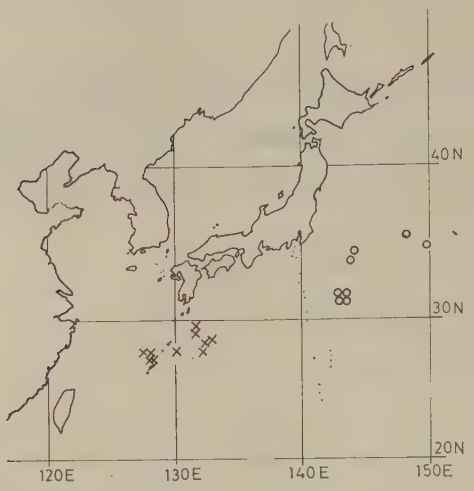

Fig. 1. Location of water-sampling site upstream and downstream along the axis of the Kuroshio current.

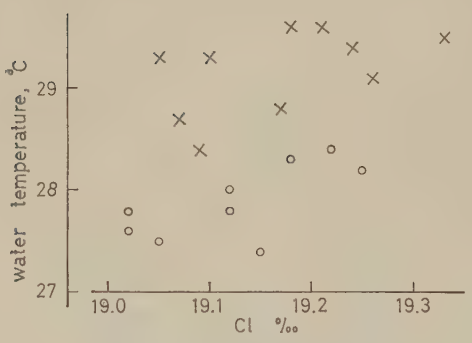

Fig. 2. Temperature and chlorinity of surface waters collected at upstream and downstream sampling sites in the Kuroshio.

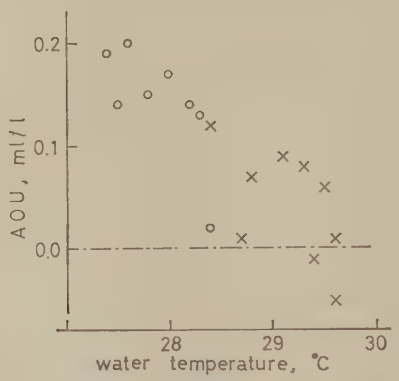

Fig. 3. An AOU vs. temperature plot for the Kuroshio surface waters.

almost coincides with that farther from it. while on the other hand, the temperature range for one of two groups of waters are distinctly separated from another. This fact suggests that the difference in water temperature is not due to mixing but to cooling of sea water at surface.

Fig. 3 shows the AOU vs. water temperature plot. According to Fig. 3, AOU is approximately 0.1 m*l*/*l* larger in the area farther from the source region than in that closer to it. Incidentally, it can be recognized by taking an example of the Japan Sea to be described later into account that the size of AOU does

not depend only on the biochemical factor such as photosynthesis or oxidative decomposition. The water temperature may be considered at the zero AOU to be one in the gas dissolution equilibrium. Fig. 3 shows that it is between 29 and 30°C.

As mentioned previously,

$$\mathrm{AOU}=O_s(t_t)-O\,,$$

which can be converted into

$$=O_s(t_h)-O-O_s(t_h)+O_s(t_t)\,,$$

where

O : the observed concentration of oxygen, ml/l

$O_s(t_t)$: the saturation amount of oxygen at the *in situ* water temperature, t_t, ml/l.

$O_s(t_h)$: the saturation amount of oxygen at a water temperature, t_h, higher than t_t, ml/l.

On the other hand,

$$O=O_a-O_o+O_p\,,$$

where

O_a: the saturation amount of oxygen in the gas dissolution equilibrium, ml/l

O_o: net oxygen depletion due to the oxidative decomposition during the whole period from the termination of dissolution equilibrium to the time of observation, ml/l

O_p: net oxygen yield due to photosynthesis during the whole period from the termination of dissolution equilibrium to the time of observation, ml/l.

Here, assuming that the water was in the dissolution equilibrium with the atmospheric oxygen at the water temperature of t_h °C but after that no more oxygen is dissolved even though the water temperature falls down to t_t °C with flow of water to the north, then

$$O_a=O_s(t_h)\,.$$

Therefore,

$$\mathrm{AOU}=O_s(t_t)-(O_s(t_h)+(O_p-O_o))\,.$$

This equation shows that the AOU vs. $O_s(t_t)$ plot gives a straight line with a slope of 1 and the $O_s(t_t)$ at the zero AOU is equal to $(O_s(t_h)+(O_p-O_o))$. The value of $(O_s(t_h)+(O_p-O_o))$ is larger when the effect of photosynthesis exceeds that of oxidative decomposition and *vice versa*.

Let us see Fig. 4 based on the consideration mentioned above. Most of

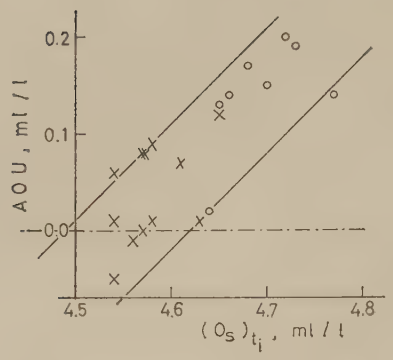

Fig. 4. A plot of AOU vs. saturation amount of oxygen at the *in situ* temperature and chlorinity for the Kuroshio surface waters.

points are distributed between two straight lines each of which has a slope of 1 and the value of $O_s(t_t)$ at the zero AOU is 4.49 or 4.63 ml/l. In Fig. 4, points nearer the straight line on the right hand side express the case in which the effect of photosynthesis exceeds that of decomposition. Points nearer the straight line on the left hand side express the case in which the effect of decomposition exceeds that of photosynthesis. Somewhere between both extremes photosynthesis balances decomposition and $O_s(t_t)$ is equal to $O_s(t_h)$. Therefore,

$$4.49 < O_s(t_h) < 4.62 \text{ m}l/l.$$

Considering the water with the chlorinity of 19.2‰ as the Kuroshio water,

$$30.2 > t_h > 28.7°\text{C}.$$

The result obtained here shows that even though the water in the dissolution equilibrium at 29 or 30°C flows north and its temperature falls by 1°C, no more oxygen dissolves into water, as the result of which AOU increases by more than 0.1 ml/l downstream.

3. Distribution of AOU in the surface waters along the axis of the Oyashio

The Oyashio is the current which reaches the region off the east coasts of Hokkaido and the northern part of Honshu through the region off the east coast of the Kamchatka Peninsula (Dodimead, Favorite and Hirano, 1963). Then, the water in the region off the east coasts of Hokkaido and the northern part of Honshu as one farther from and the water in the Bering Sea as one closer to the source region of the current are chosen.

Figs. 5 and 6 show the AOU vs. t_t and the AOU vs. $O_s(t_t)$ diagrams pertaining to surface waters in those regions. In the similar way as mentioned previously,

$$6.79 < O_s(t_t) < 7.03 \text{ m}l/l,$$

where $O_s(t_t)$ shows the saturation amount of oxygen at the water temperature of t_t °C which is somewhat lower than the *in situ* water temperature of t_t °C. The result obtained here shows that even though the water in the dissolution equilibrium at about 7°C flows south and the water temperature rises by 4°C, oxygen

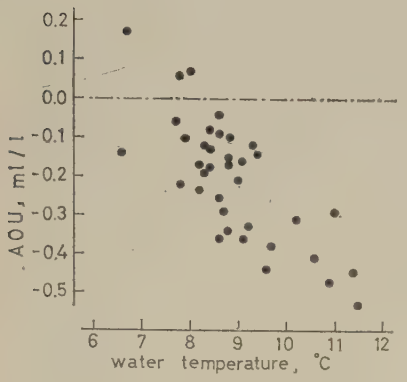

Fig. 5. An AOU vs. temperature plot for the Oyashio surface waters.

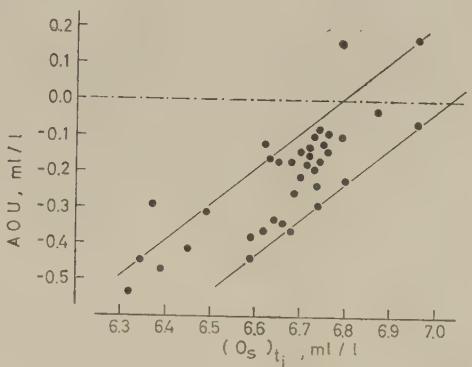

Fig. 6. A plot of AOU vs. saturation amount of oxygen at the *in situ* temperature and chlorinity for the Oyashio surface waters.

in the water does not escape out to the air, as the result of which the value of AOU decreases by about 0.4 m*l*/*l*.

4. Distribution of AOU in the surface waters of the boundary region between Kuroshio and Oyashio

The region off the east coast of Honshu is the region where the end of the Kuroshio meets the end of the Oyashio. The representative values of water temperature, chlorinity and oxygen amount are 10.0°C, 18.1‰, 6.80 m*l*/*l*; 27.0°C, 19.2‰, 4.55 m*l*/*l* respectively at the ends of the Oyashio and the Kuroshio. Now, assuming that both species of waters mix with each other at various ratios in the boundary region and the oxygen concentration of 4.55 or 6.80 m*l*/*l* is kept

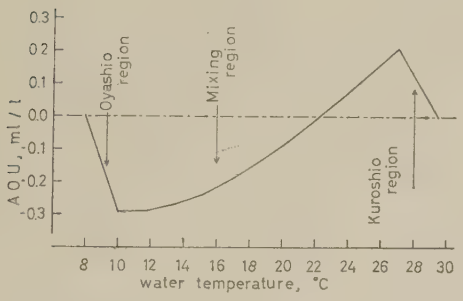

Fig. 7. The calculated curve expressing the relationship between AOU and water temperature in the region of the Kuroshio and the Oyashio.

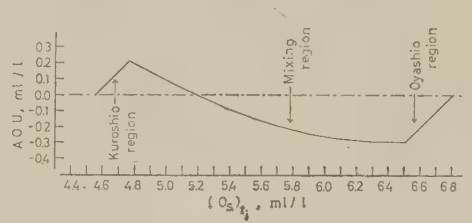

Fig. 8. The calculated curve expressing the relationship between AOU and oxygen saturation amount at the *in situ* temperature and chlorinity in the regions of the Kuroshio and the Oyashio.

unchanged in the main region of each current since the time of gas dissolution equilibrium, then the curves representing the relationship between AOU vs. t_i and that between AOU vs. $O_s(t_i)$ were obtained by calculation as shown in Figs. 7 and 8. Observational results are shown in Figs. 9 and 10. As seen in these figures, the calculated curves fit fairly well with the observational results.

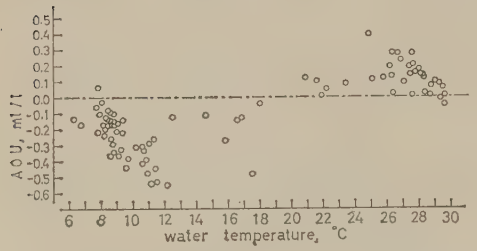

Fig. 9. The observed relationship between AOU and water temperature in the regions of the Kuroshio and the Oyasio.

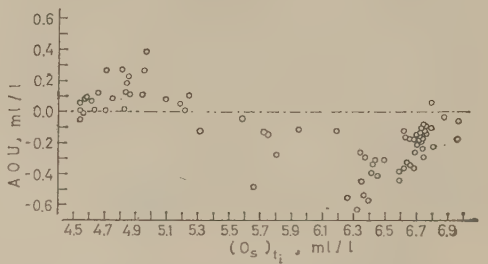

Fig. 10. The observed relationship between AOU and saturation amount of oxygen at the *in situ* water temperature and chlorinity in the regions of the Kuroshio.

5. Seasonal variation of AOU in the surface waters of Kuroshio

In the source region of the Kuroshio (around lat. 15°N), the surface water temperature is about 29°C in summer and about 25°C in winter. The Kuroshio water flows north. At lat. about 35°N, the surface water temperature is about 25°C in summer and about 13°C in winter. Accordingly, the difference of water temperature between at lat. 35°N and at lat. 15°N is about 4°C in summer and about 12°C in winter. If the concentration of oxygen dissolved at higher water temperature is kept unchanged even though water flows north, the value of AOU seems to become larger in winter than in summer. The trend expected here has been confirmed in the observational result as shown in Fig. 11.

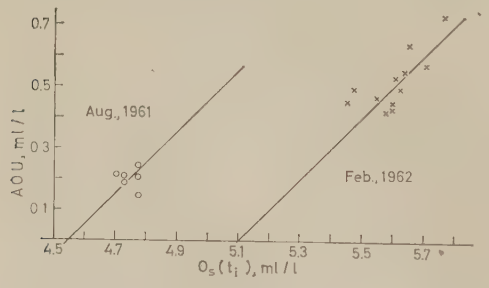

Fig. 11. The relationship between AOU and saturation amount of oxygen at the *in situ* temperature and chlorinity for the Kuroshio waters in summer and in winter.

Fig. 11 shows the observational result obtained aboard R/V the *Ryofu Maru* in August 1961 and in February

1962. According to Fig. 11, AOU in February is about 0.3 ml/l larger than in August. In this figure, the intersection of a straight line with a slope of 1 passing the center of scattering of points and the axis of AOU=0 gives about 4.55 ml/l as the value of $O_s(t_i)$ in the case of August. Then, assuming that $O_s(t_i)$ is nearly equal to $O_s(t_h)$ and that the chlorinity is 19.2‰, the value of t_h becomes 29.5°C. Similarly, in the case of February, the value of t_h becomes 23.0°C. It is noteworthy that those values of t_h in summer and in winter are approximately equal to the surface water temperatures in both seasons in the source region of the Kuroshio.

6. Distribution of AOU in the surface waters of the Japan Sea

Assuming that the whole of the surface waters of the Japan Sea is originated from the warm water of the Tsushima Current, a branch of the Kuroshio, it may be expected that AOU in the surface waters increases downstream, in other words, it increases with the decrease of water temperature. But, actually, it is not the case. AOU decreases with water temperature as shown in Fig. 12.

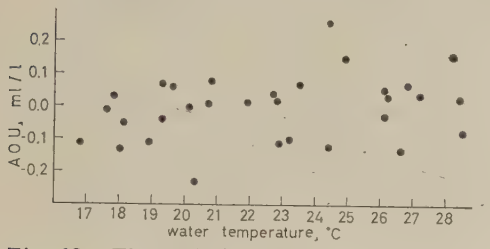

Fig. 12. The relationship between AOU and temperature of waters in the Japan Sea.

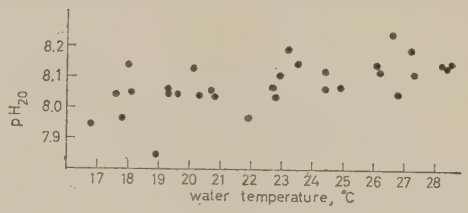

Fig. 13. The relationship between pH at 20°C and temperature of waters in the Japan Sea.

Fig. 13 shows the relationship between pH at 20°C and water temperature of surface waters of the Japan Sea. pH is lower in the water with lower temperature. If the decrease of AOU is caused by photosynthesis, pH must increase with the decrease of water temperature, which is the reverse of the fact. Consequently, it may be concluded that the main factor governing the phenomenon is not the photosynthesis.

The similarity of the relationship between the AOU and water temperature in the Japan Sea to that in the boundary region between Kuroshio and Oyashio suggests that it is mixing of both kinds of waters, one with higher water temperature and higher positive AOU and another with lower water temperature and lower negative AOU.

7. Conclusion

Surface water is not exactly saturated with atmospheric oxygen at the *in situ* water temperature and chlorinity, the cause of which has so far been ascribed to the biochemical effect such as photosynthesis or oxidative decomposition of organic material. But there seems to be another cause. The deviation, positive or negative, from the saturation amount is larger in its absolute value at the end of the current in the case of either Kuroshio or Oyashio. This may be thought to be an indication of the slower rate of gas exchange between sea and air as compared to the rate of heat exchange between sea and air as compared to the rate of heat exchange. The slower rate of gas exchange may also be recognized in the seasonal variation of AOU.

This conclusion is only tentative, because it is still uncertain whether or not the Fox's saturation amount is suitable enough and the precision of the Winkler method routinely in use is durable enough to be employed for such kind of consideration.

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Distribution of pH in the Kuroshio and the Oyashio regions and its oceanographical significance

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Abstract: The calculated curve based on pH at 20°C in place of *in situ* pH and carbonate alkalinity fits well with the observed result. The pH consists of two parts: one dependent on temperature and chlorinity and another dependent on AOU. The former allows us to evaluate the partial pressure of atmospheric CO_2 above the sea surface. The evaluated partial pressure of CO_2 was too low in the Kuroshio surface waters and too high in the Oyashio surface waters, as far as the downstream part of the respective current is concerned. The cause arising the above-mentioned trend may be explained in the following way: The rate of gas exchange is much slower than is the rate of heat exchange between air and sea.

As to the deep waters, in the Pacific the evaluated partial pressure of CO_2 was also too low, which may be ascribed to the dissolution of calcium carbonate in the deep. But, in the Japan Sea, the evaluated partial pressure of CO_2 is nearly equal to the normal value expected under one atmosphere of air, which suggests that waters of the Japan Sea are rather poor in the calcium carbonate content.

1. Introduction

There are few papers dealing with pH of sea water particularly from the chemico-oceanographical view-point. SARUHASHI (1956) calculated the vertical distribution of H_2CO_3, HCO_3^- and CO_3^{2-} based on the data by HAMM and THOMPSON. PARK (1966) pointed ont the existence of pH maximum in the deep and tried its chemico-oceanographical interpretation. Recently, AKIYAMA, SAGI, YURA and MAEDA (1968) clarified the distributions of *in situ* pH and total alkalinity in the western North Pacific.

In this paper, the present anthor employed data obtained by chemists of the Japan Meteorological Agency aboard R. V. *Ryofu Maru* in the frontal region off the north-eastern coast of Honshu, which includes the downstream region of the Kuroshio and that of the Oyashio and the boundary of both. He has proposed

to adopt a new treatment of pH by which the partial pressure of atmospheric CO_2 above the sea surface can be evaluated. Then, the peculiarities in the evaluated partial pressure of atmospheric CO_2 have been clarified for waters occupying the downstream regions of both currents and the deep waters in the western North Pacific and the Japan Sea. The chemico-oceanographical interpretation is attempted.

2. The relationship between pH and AOU

Fig. 1 shows the relationship between pH and AOU. The ordinate in Fig. 1 expresses pH at 20°C which is denoted by pH_{20}. The pH_{20} can be evaluated by the following equation:

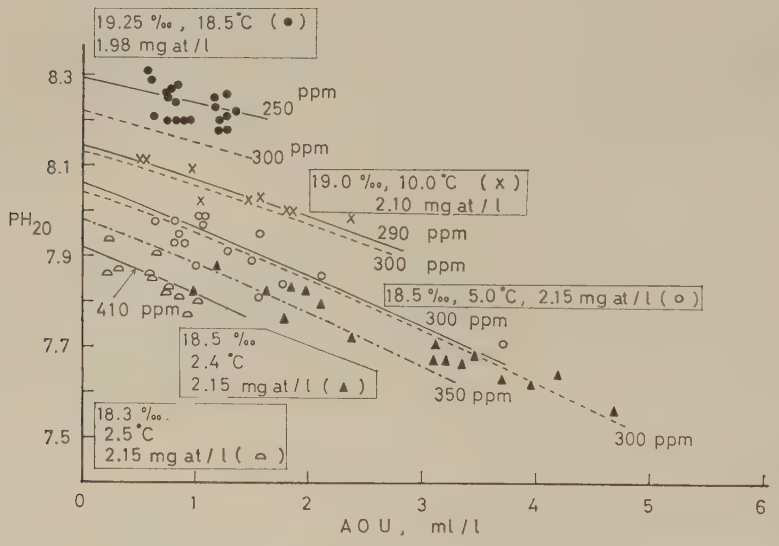

Fig. 1. The pH vs. AOU diagram for the Kuroshio and Oyashio waters.

$$pH_{20}=pH_t+\alpha(20-t)\,.$$

The coefficient α is about -0.01. The purpose of use of pH_{20} is to make clearer the relationship between pH and CO_2 by canceling the temperature effect on pH out.

A group of points distributed at the uppermost part of Fig. 1 express the Kuroshio waters with temperature of 18.5°C and chlorinity of 19.25‰. They occupy the depths ranging from 50 to 200 m. A group of points distributed at the lowest part of Fig. 1 express the Oyashio waters with temperature of 2.5°C

and chlorinity of 18.3‰. They occupy the depths ranging from 50 to 100 m.

It may be seen in Fig. 1 that pH of sea water consists of two parts: one dependent on water temperature and chlorinity and another dependent on AOU. Let us call the former the conservative pH and denote it as pH_c. Then, results of Fig. 1 indicate that waters with higher temperature and chlorinity have higher pH_c and *vice versa*. Since the solubility of CO_2 in sea water is lower at higher temperature and chlorinity, the above-mentioned results support the idea that pH_c depends primarily on the amount of CO_2 dissolved from the atmosphere. But, here it must be noticed that even though the amount of CO_2 is the same, the decrease of pH is more evident for lower salinity. Therefore, pH should be considered in connection with alkalinity.

The variable component of pH depending on AOU is called non-conservative pH and denoted by ΔpH. ΔpH is negative and its absolute value increases with AOU, as shown in Fig. 1. The increase of AOU is equivalent to the increase of free CO_2 produced by the oxidative decomposition of organic material. Assuming no variation of alkalinity at the oxidative decomposition of organic material, the relationship between AOU and pH can be derived in the following way:

According to Buch's study (Harvey, 1957),

$$C = f_1 A \qquad (1)$$

$$P = f_2 A \qquad (2)$$

$$f_1 = (1+(K_2'/a_H{}^{\cdot}) + (\alpha_s a_H{}^{\cdot}/K_1'\alpha_o a_{H_2O})/(1+(2K_2/a_H{}^{\cdot}))$$

$$f_2 = a_H{}^{\cdot}/K_1'\alpha_0(1+(2K_2/a_H{}^{\cdot})\, a_{H_2O}$$

where

C : the amount of total carbon dioxide, m-moles/l

A : carbonate alkalinity, m-eq/l

P : partial pressure of CO_2, 10^{-3} atm

K_1', K_2' : first and second apparent equilibrium constant of carbonic acid in sea water, respectively

α_0 : solubility of CO_2 in distilled water, moles/l

α_s : solubility of CO_2 in sea water, moles/l

a_{H_2O} : activity of water

a_H : activity of hydrogen ion, g ion/l.

At a certain temperature and chlorinity, f_1 and f_2 are functions of solely pH. The decrease of pH corresponds to the increase of f_1 and f_2. Denoting the amount of total carbon dioxide at the zero AOU as C_e and f_1 corresponding to pH_c as f_{1c},

$$C_c = f_{1c}A\,. \qquad (3)$$

From equations (1) and (3),

$$C-C_c=(f_1-f_{1c})\,A=(f_1-f_{1c})(C_c/f_{1c})\,. \qquad (4)$$

According to SUGIURA and YOSHIMURA's study on the total carbon dioxide (1967),

$$C-C_c = (0.1/3)\,\mathrm{AOU}\,. \qquad (5)$$

From equations (4) and (5),

$$f_1 = f_{1c} + \mathrm{AOU}/(30\,C_c/f_{1c}) \qquad (6)$$

is obtained.

The value of C_c is known from the C-AOU diagram which was obtained by SUGIURA and YOSHIMURA for each group of waters. Now, let us consider about waters with the particular temperature and chlorinity concerned. A value of pH is arbitrarily chosen. Substituting the value of f_{1c} corresponding to pH_c into equation (6), a relationship between f_1 and AOU is given. For waters with known water temperature and chlorinity, f_1 determines pH. Consequently, equation (6) gives the relationship between pH and AOU. When the value of pH_c is changed, another relationship can be obtained. In this way, several curves expressing relationships between pH and AOU can be obtained for various values of pH_c.

If the above-mentioned consideration is wrong, the relationship between pH and AOU coincident with the observational result may not be expected. But, actually we could find out the relationship which fits well with the observation for an appropriate pH_c, as shown in Fig. 1. pH_c determines f_{2c}, which in turn determines P_c from equation (2). P_c is the partial pressure of CO_2 in water saturated with air at the *in situ* temperature and chlorinity under one atmosphere. It is known to be 300 to 320×10^{-4} atm. But, P_c evaluated from the results of Fig. 1 is not all coincident with the partial pressure of atmospheric CO_2 under one atmosphere. It is quite low like 230×10^{-4} atm in the Kuroshio waters and quite high like 360×10^{-4} atm in the Oyashio waters. Is this opposite trend just accidental?

3. pH of the surface waters

It may be interpreted in the following way that P_c is too low in the Kuroshio surface waters but too high in the Oyashio surface waters: The Kuroshio waters are saturated with atmospheric CO_2 and O_2 at higher temperature in the source region. Even though they flow to the north and the water temperature falls due to the release of heat energy from the surface, the amount of dissolved gas is

kept unchanged. This assumption brings out the depression of the partial pressure of CO_2. The similar discussion may be possible in the case of the Oyashio surface waters, too.

In the following computation, it is assumed that the rate of gas exchange between air and sea is very slow but the dissociation of carbonic acid in sea water proceeds without any delay at the *in situ* water temperature and chlorinity. Practically, we assume that the amount of gas dissolved at 23°C does not change after the water temperature falls down to 18.5°C in the downstream region of the Kuroshio. But, the dissociation equilibrium of carbonic acid occurs at 18.5°C which is the *in situ* water temperature. In other words, the value at 23.0°C is used as α_0 and α_s, and the value at 18.5°C as f_1 and f_2. In addition, when water temperature rises from 18.5 to 23.0°C, the saturation amount of oxygen decreases by 0.40 m*l*/*l*. Therefore, under the assumption of no change in the amount of dissolved gas in spite of fall of water temperature after flow to the north, all points in the pH-AOU diagram should be shifted to the left by 0.04 m*l*/*l* in parallel with the axis of abscissa (see Fig. 2). Consequently, 290×10^{-4} atm is obtained as the value of P_c, which is closer to the normal partial pressure of atmospheric CO_2 under one atmosphere.

Fig. 3 shows the relationship for the Oyashio surface waters. In this case, assuming that even though the temperature of water saturated at 0°C with air under one atmosphere rises to 2.5°C, the amount of dissolved gas is kept unchanged, 290×10^{-4} atm is obtained as the value of P_c. It may be seen by these two examples that when ocean current runs in the direction of north to south after a long run in the direction of east to west, there is a possibility of appearance of the difference of the partial pressure of CO_2 between the calculated value

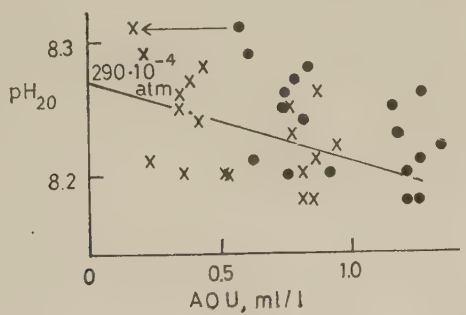

Fig. 2. The pH vs. AOU diagram for the Kuroshio waters at the depth of 50 to 200 m.

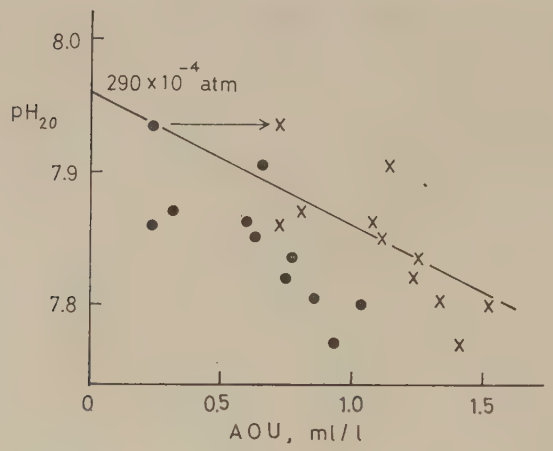

Fig. 3. The pH vs. AOU diagram for the Oyashio waters at the depth of 50 to 100 m.

based on pH_c and the normal value in the air above the sea surface under one atmosphere. It is because of the delay of gas exchange in comparison with heat exchange between air and sea.

4. pH of the deep waters

Fig. 4 shows the relationship between pH_{20} and AOU as to the deep waters of the Pacific and the Japan Sea. In the case of the Pacific, temperature and

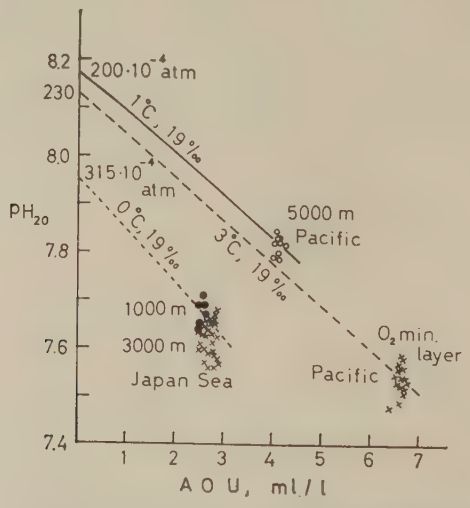

Fig. 4. The pH vs. AOU diagram for the Kuroshio and the Oyashio surface waters.

chlorinity at the depth of 5,000 m are about 1°C and 19.0‰ respectively. As the value of P_c 200×10^{-4} atm is obtained. For the oxygen minimum layer ranging from 850 to 1,350 m in the depth, 230×10^{-4} atm is obtained as the value of P_c. These values are either too low as compared with the normal partial pressure of atmospheric CO_2 under one atmosphere. If it is true that there is the dissolution of $CaCO_3$ in the deep, the partial pressure of CO_2 may be depressed by the following reaction:

$$CO_2 + H_2O + CaCO_3 = Ca(HCO_3).$$

Therefore, the value of P_c obtained here for the Pacific deep waters seems to be quite reasonable under the assumption of the dissolution of $CaCO_3$ in the deep. But, on the contrary, in the case of the Japan Sea, 315×10^{-4} atm was obtained for waters at depths of 1,000 and 3,000 m. There is no difference between the calculated value and the normal partial pressure of atmospheric CO_2. This result seems to suggest that the calcium carbonate content is probably low in the waters of the Japan Sea.

5. Conclusion

Although several workers (Sillen, 1961; Garrels, 1965; Mackenzie and Garrels, 1965) give a stress on the role of silicate minerals upon the maintenance of sea water pH near 8, the features in the distribution of pH in the sea area around Japan can be explained primarily by reactions occurring in the oceanic carbon dioxide system.

In the downstream region of the Kuroshio, the observed pH of the surface waters is too high as compared to the pH evaluated based on the assumption of the dissolution equilibrium of atmospheric CO_2 at the *in situ* water temperature and chlorinity under one atmosphere. This trend can be explained by assuming that the amount of CO_2 dissolved at higher temperature in the source region is almost kept unchanged even after the flow to the north, in other words, the delay of gas exchange in spite of rather quick heat exchange between air and sea. The similar consideration can be applied to the pH of the Oyashio waters.

In the deep waters of the eastern North Pacific, the observed pH is too high as compared with the pH evaluated based on the assumption of the dissolution equilibrium of atmospheric CO_2 at the *in situ* water temperature and chlorinity. This trend seems to be ascribed to the disappearance of free CO_2 by the dissolution of $CaCO_3$ in the deep. It seems to suggest the meager content of $CaCO_3$ in

the waters of the Japan Sea that P_c evaluated in the Japan Sea from pH_c takes a value of 315×10^{-4} atm.

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The relationship between phosphate and oxygen concentrations in the up-welling waters and its oceanographical significance

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Abstract: The concentrations of phosphate and oxygen were measured in the up-welling region off the Oregon Coast. As the relationship between the concentrations of both elements were obtained several straight lines, each of which had the slope equivalent to the oxidative ratio of the marine plankton. The interuption of straight lines is caused by the displacement of water mass. The depth from which water rises might occassionally change.

1. Introduction

The up-welling water comes up from the subsurface depth of several hundred meters. Since the rising-up waters might be of the same water mass, in other words, they might have the same concentration of conservative phosphate, the relationship between phosphate and oxygen concentrations might be characteristic of photosynthesis as to waters collected at several depths in the euphotic zone. The present author wanted to ascertain this idea.

Fortunately, he had an opportunity to measure the concentrations of phosphate and oxygen in the up-welling region along the Oregon Coast. Observation was done aboard R. V. *Yaquina* in co-operation with K. PARK, L. GORDON and M. CISSELL, Oregon State University for 10 days since July 5, 1967. The present paper describes the theoretically expected relation first and then the result of observation. Its oceanographical significance is discussed.

2. The expected relationship between phosphate and oxygen concentrations for the surface waters of the up-welling region

When the up-welling water is rising from the subsurface layer to the euphotic layer, the phosphate concentration decreases and the oxygen concentration increases with rising of water because of the predominance of photosynthesis over

oxidative decomposition of marine plankton. Let us denote the concentrations of phosphate and oxygen of water collected at the depth of u and at time of t respectively as P_u and O_u. And let us denote the concentrations of phosphate and oxygen of water collected at the depth of i, which is shallower than u, and at the time of t respectively as P_i and O_i. Then,

$$P_i = P_u' - \Delta P, \qquad O_i = O_u' + \Delta O$$

where P_u' and O_u' show respectively concentrations of phosphate and oxygen at the time of $(t-\Delta t)$ when the water occupying the depth of i at the time of t has passed the depth of u. ΔP and ΔO show increments in concentrations of phosphate consumed and oxygen produced respectively due to the effect of photosynthesis during the water rises from the depth of u to the depth of i for the period of Δt. Then, the differences between P_u and P_i and between O_i and O_u are expressed as follows:

$$P_u - P_i = (P_u - P_u') + \Delta P \qquad (1)$$

$$O_i - O_u = (O_u' - O_u) + \Delta O. \qquad (2)$$

If $P_u = P_u'$ and $O_u = O_u'$,

$$P_u - P_i = \Delta P \qquad (3)$$

$$O_i - O_u = \Delta O. \qquad (4)$$

Equations (3) and (4) suggest that the $(P_u - P_i)$ vs. $(O_i - O_u)$ plot is expected to give the particular relationship between the consumed phosphate and produced oxygen at photosynthesis.

When $P_u \neq P_u'$ and $O_u \neq O_u'$,

$$(P_u - P_i) - (P_u - P_u') = \Delta P \qquad (5)$$

$$(O_i - O_u) - (O_u' - O_u) = \Delta O. \qquad (6)$$

Equations (5) and (6) suggest that when the origin of the co-ordinate system is shifted by $(P_u - P_u')$ and $(O_u' - O_u)$ with kept both axes of co-ordinate in parallel with the original ones, the relationship between phosphate and oxygen shown by equations (5) and (6) coincides exactly with that shown by equations (3) and (4). The shift of the origin is caused by the non-biological factor, such as the displacement of water mass in Δt.

As mentioned above, the $(P_u - P_i)$ vs. $(O_i - O_u)$ plot generally makes a curve expressing the particular biological relationship. But, when the displacement of water mass occurs, the particular biological relationship is obscured. And, again as long as waters of the same mass are rising ceaselessly, the previously mentioned particular relationship can be seen. Consequently, it is expectable that the $(P_u - P_i)$ vs. $(O_i - O_u)$ plot gives, in general, several parallel curves.

3. The observed relationshjp between phsphate and oxygen concentrations of the surface waters in the up-welling region

Fig. 1 shows the observed relationship between (P_u-P_i) and (O_i-O_u). According to Fig. 1, it can be seen that several points expressing waters collected

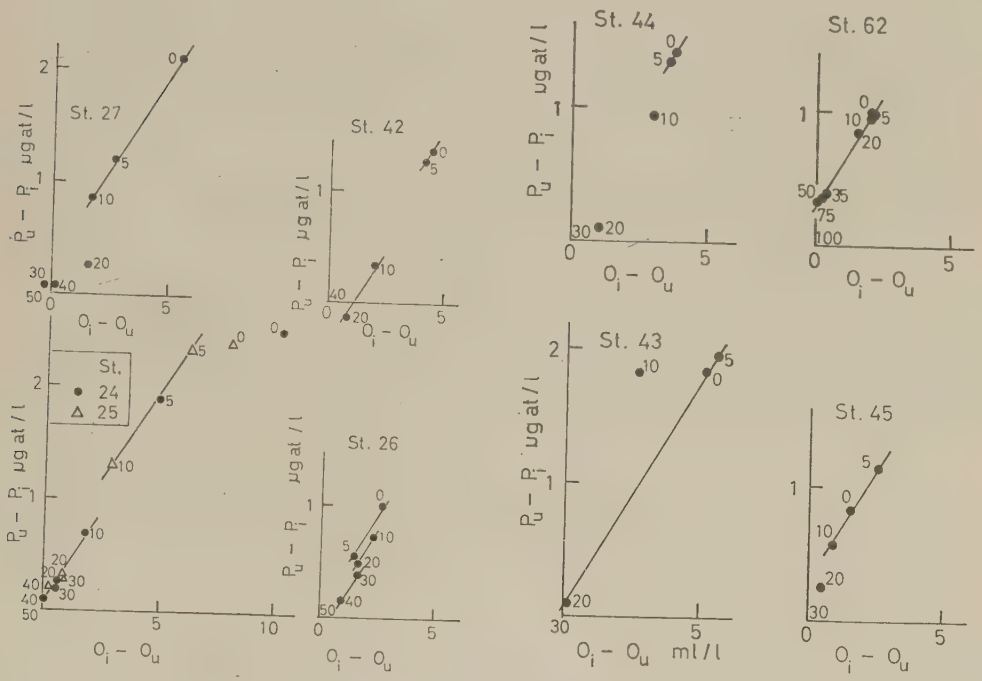

Fig. 1. The relationship between (P_u-P_i) and (O_i-O_u). Figures aside plotted points or the origin of co-ordinate show the depth of water collection.

at the neighbouring depths are located on a straight line whose slope is equivalent to 1 µg at P/l over 3 ml O_2/l. This slope is equivalent to the oxidative ratio of marine plankton with the average composition (SUGIURA and YOSHIMURA, 1964). It is also coincident with the slope found outside the up-welling region off the Oregon Coast (see Fig. 2).

At stations 26 and 42 shown in Fig. 1, two or three parallel straight lines can be seen. And waters represented by two points on each straight line occupy the nearest depth, for example, 0 and 5, 10 and 20, and 30 and 40 m. Remarkable in particular among nine examples shown in Fig. 1 is an example of station 62. In this case, waters occupying depths of 0, 5, 10, 20, 35, 50 and 75 m may be

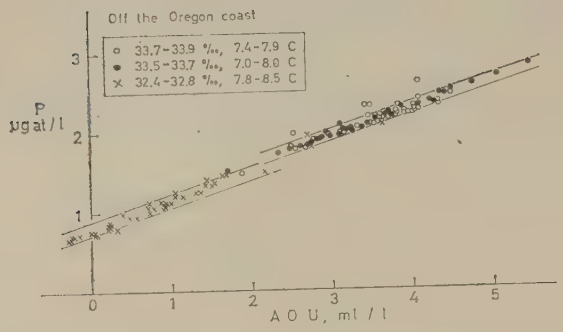

Fig. 2. The relationship between phosphate and AOU outside the up-welling region off the Oregon Coast.

thought to be of the same mass. Between 75 and 100 m, the displacement of water mass occurred. When $(P_{i+1}-P_i)$ $(=\Delta P)$ and (O_i-O_{i+1}) $(=\Delta O)$ are respectively plotted against the depth, the relationship shown in Fig. 3 is obtained, where the depth of i is shallower than the depth of $(i+1)$. This can be thought to express the vertical distribution of the rate of photosynthesis *in situ*. From the pattern of its vertical distribution, it is inferred that the concentrations of phosphate and oxygen expectable at 100 m depth of the water which occupied the

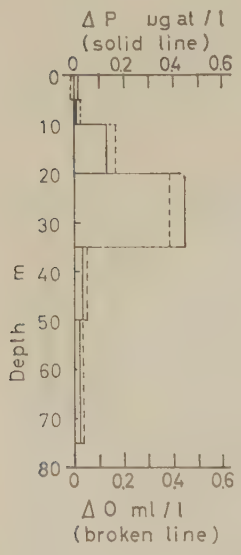

Fig. 3. The vertical distribution of ΔP and ΔO at Sta. 62.

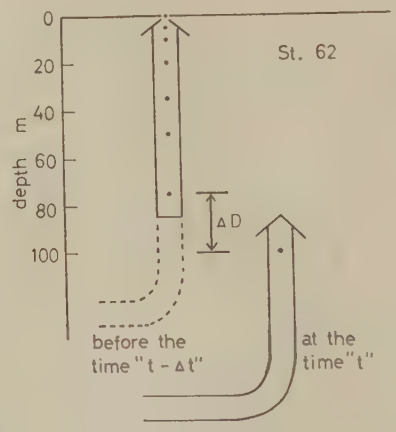

Fig. 4. Illustration of displacement of water mass in the up-welling water system.

depths of 0 to 75 m at the time of observation might be quite close to those at 75 m depth. Accordingly, the length of shift of the origin is equivalent to less than 0.3 μg at/l in phosphate and less than 0.1 ml/l in oxygen.

That is, $P_n - P_o \leq 0.3$ μg at/l and $O_o - O_n \leq 0.1$ ml/l, where the subscript n and o mean waters which occupied the depth of 100 m respectively before the time of $(t-\Delta t)$ and at the time of t.

Since the phosphate concentration generally increases and the oxygen concentration decreases with the depth, the result obtained at station 62 suggests that water occupying the depth of 100 m before the time $(t-\Delta t)$ came up from the depth shallower than the depth from which the water occupying the same depth at the time of observation t came up. Waters which occupied depths of 0, 5, 10, 20, 35, 50 and 75 m are all of the same origin as is water which occupied the depth of 100 m before the time $(t-\Delta t)$. The relation mentioned here is illustrated in Fig. 4.

$$\Delta t \leq \Delta D/v\,,$$

where D means 25 m in this case and v shows the upward velocity of the upwelling waters.

4. Conclusion

Concentrations of phosphate and oxygen of water at a certain depth are chosen as the references. The reference depth ranges from 30 to 100 m, depending on the station, but it is chosen arbitrarily.

When the difference between the phosphate concentration at the reference depth and that at the depth shallower than the reference depth is plotted against the difference between the oxygen concentration at the depth shallower than the reference depth and that at the reference depth, several straight lines are found, each of which connects several points representing waters collected at the neighbouring depths and its slope is equivalent to 1 μg at P/l over 3 ml O_2/l, which is equal to the oxidative ratio applied to the marine plankton (Sugiura and Yoshimura, 1964).

Interuption of straight lines may be interpretted as the displacement of water mass. In the case of the example at station 62, the boundary between the neighbouring water masses was located somewhere between depths of 75 and 100 m. The concentrations of phosphate and oxygen suggest that water occupying the depths between 0 and 75 m at the time of observation came up from the depth shallower than the depth from which the water occupying the depth of

100 m at the same time came up. Therefore, there might be a possibility of the pulsation of upwelling accompanying the occassional variation of the depth from which the water rises.

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Carbon dioxide in the atmosphere and in sea water in the adjacent seas of Japan

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Abstract: In 1968, the measurements of partial pressures of carbon dioxide in the atmosphere (pco_2) and in sea water (Pco_2) were carried out in regions southeast of Japan and in the East China Sea. In the Pacific Ocean southeast of Japan, the pco_2 ranges between 318 and 332 ppm, and the Pco_2 shows wider variations locally ranging from 272 to 519 ppm. Generally the pco_2 increases and in contrast with that the Pco_2 decreases with increasing latitudes, while the longitudinal variations are almost negligible. The striking high Pco_2 are observed in a belt area from 22.5°N to 25°N corresponding with the Subtropical Countercurrent where the conspicuous deepening of the isotherms is characteristic. In the East China Sea the pco_2 distributes relatively homogeneous, being around 323 ppm. The Pco_2 varies locally in the range between 295 and 372 ppm, being lowest in the Kuroshio region and highest in the cold water region of the Yellow Sea. Vertically the Pco_2 shows the almost constant values at the surface layer shallower than the depth of thermocline, and below the depth it increases gradually in the Kuroshio region and sharply in the cold water region.

1. Introduction

It has been well known that the partial pressure of carbon dioxide is fairly constant in the atmosphere while it varies considerably from area to area in the surface ocean water and the equilibrium of carbon dioxide between the atmosphere and the surface ocean water is not achieved locally (Takahashi, 1961; Waterman, 1964; Miyake and Sugimura, 1968; Keeling, 1968). The result of measurements of carbon dioxide in the western North Pacific Ocean by the present author was reported in the previous paper (Akiyama, 1968). In this paper, the result of study on the partial pressure of carbon dioxide in the atmosphere (pco_2) and in sea water (Pco_2) using an infrared absorption method in the adjacent seas of Japan is reported. The measurements were carried out on board the R. V. *Ryofu Maru* in the Pacific Ocean southeast of Japan in May, and in the East China Sea in October, 1968. In addition to the problems of the long-term variations in the

atmosphere and the transfer through the air-sea interface of carbon dioxide, the correlation between the Pco_2 and other hydrographic elements and the vertical variation of the Pco_2 in the ocean are studied in the present work.

2. Method of measurement

The partial pressure of carbon dioxide in the atmosphere was measured continuously by the infrared analyser (Beckman-Tōshiba Model IR-315). Water vapor was removed before the air sample was analysed by passing through a cold trap (Hayakawa Denki Model TE-33) at the temperature of 1.5°C.

The partial pressure of carbon dioxide in sea water was measured by the equilibrator method reported in the previous paper (Akiyama, 1968).

3. Results and discussions

1. The Pacific Ocean southeast of Japan

The pco_2 and the Pco_2 in the Pacific Ocean southeast of Japan are shown

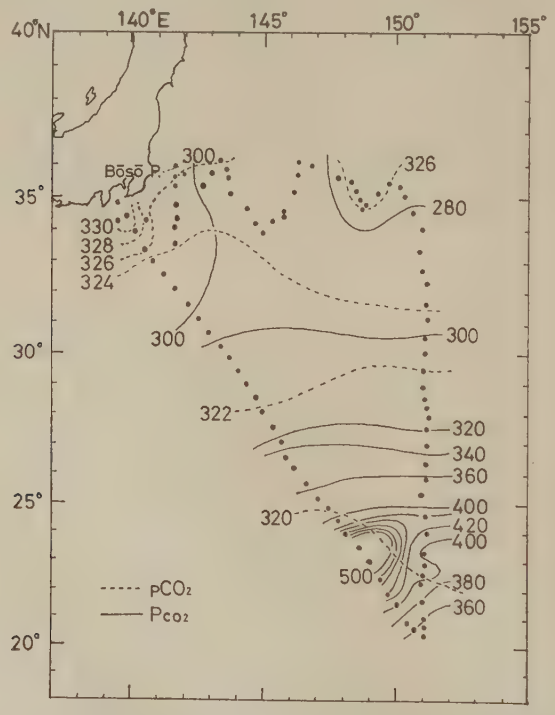

Fig. 1. Horizontal distributions of the partial pressures of carbon dioxide in the atmosphere (pco_2) and in surface water (Pco_2) in the sea southeast of Japan.

in Fig. 1. The pco_2 was measured continuously and the Pco_2 every 3 hours during the cruise.

The pco_2 ranges between 318 and 332 ppm in the observed area. Higher values than 326 ppm are observed in the coastal area southeast of Bōsō Peninsula, which may be caused by the influence from the land where much carbon dioxide is released to the atmosphere. Lower values less than 320 ppm are observed in the southern area along 20°N. A rate of variation in the pco_2 along the meridian of 151°E is roughly 1 ppm per 150 miles on an average. Generally the variations of the pco_2 are significant along the longitude and smaller along the latitude, which may correlate with the general circulation of the air.

The Pco_2 shows wider variations ranging from 270 to 519 ppm. In Fig. 2 the latitudinal variations of the pco_2, the Pco_2 and the surface water temperature,

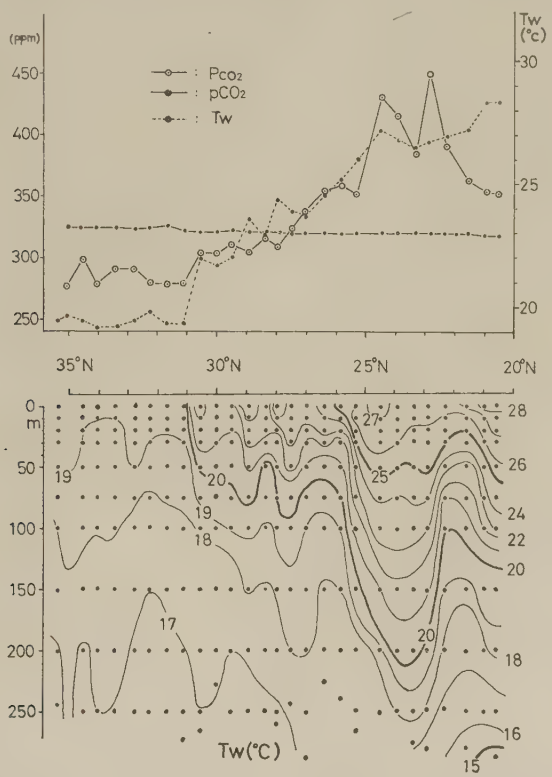

Fig. 2. Variations of the partial pressures of carbon dioxide in the atmosphere (pco_2) and in surface water (Pco_2) and surface water temperature (upper), and the vertical profile of water temperature (lower) along the meridian of 151°E.

and the vertical profile of the water temperature are shown along the meridian of 151°E. In contrast with the gradual decrease of the pco_2, the remarkable increase of the Pco_2 toward the south is observed. Sudden rises of the Pco_2 and the surface water temperature occur almost simultaneously. However, there are some exceptions as shown in the area south of 22.5°N where the Pco_2 decreases abruptly while the surface water temperature rises gradually. Along the meridian of 151°E the Pco_2 rises sharply southward at locations of 31°N, 28°N and 25°N and decreases at 22.5°N. The correlation coefficient between the Pco_2 and the water temperature is 0.73.

The striking higher values of the Pco_2 than 400 ppm is observed in a band area from 22.5°N to 25°N where the conspicuous deepening of the isotherms occurs. Recently Yoshida and Kidokoro (1967) suggested the existence of the eastward current in this zone and named it the Subtropical Countercurrent. The observed high Pco_2 may be one of the characteristics of the current originated from the subtropical western Pacific Ocean. Generally the high Pco_2 in surface water is expected when rapid warming of the water occurs and when the carbonic acid substances are supplied to the surface from the deep by upwelling and when the regeneration of carbon dioxide in decomposition processes of dead organisms occurs at relatively shallower layer. A dominant factor which causes the high Pco_2 in the Subtropical Countercurrent is not known. Miyake and Sugimura (1968) reported that there is a higher correlation between the Pco_2 in surface water and the depth of pycnocline. The lowest Pco_2 in the present work (270 ppm) is found in the Kuroshio extension. While flowing eastward, the Kuroshio water loses gradually the carbon dioxide by about 30 ppm from 140°E to 151°E. In the cross sections of the Kuroshio the difference of the Pco_2 between the south and the north edges of the Kuroshio is not found despite the larger difference in the water temperature.

The equilibrium of carbon dioxide between the atmosphere and the ocean surface is attained only in narrow band area of 27°N—28°N. In most part of the observed area the equilibrium is not achieved. In the north of the above equilibrated area the Pco_2 is less than the pco_2, and on the contrary the Pco_2 exceeds much the pco_2 in the south. The maximal differences (Pco_2—pco_2) reach −60 ppm in the Kuroshio water and +190 ppm in the Subtropical Countercurrent.

2. The East China Sea

The pco_2 observed in the East China Sea ranges from 320 to 328 ppm. Most

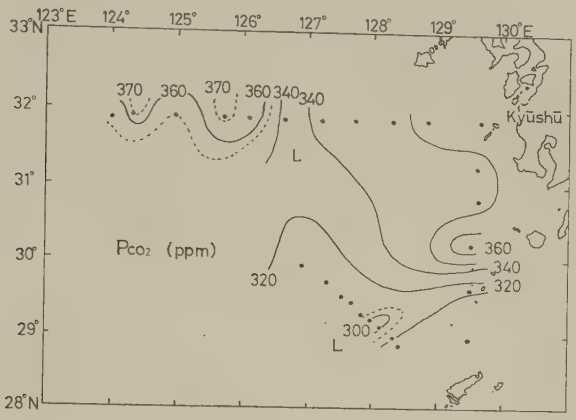

Fig. 3. Horizontal distribution of the partial pressure of carbon dioxide in surface water (Pco_2) in the East China Sea.

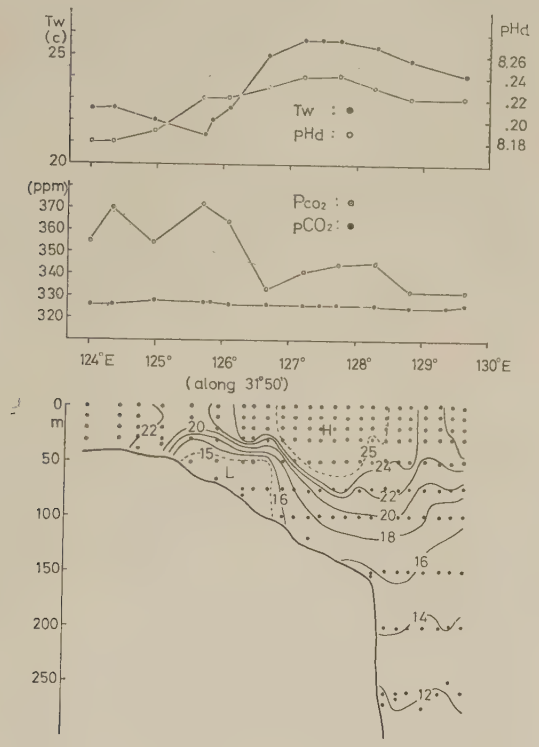

Fig. 4. Variations of the partial pressures of carbon dioxide in the atmosphere (pco_2) and in surface water (Pco_2), and surface water temperature and pH *in situ* (upper), and the vertical profile of the water temperature (lower) along the latitude of 31°50′ N.

of the pco_2 values are close to the average value of 323 ppm. Any systematic tendency in geographical distribution is not found.

The Pco_2 varies locally in the ranges between 295 and 372 ppm as shown in Fig. 3, reflecting the presence of different water masses. Along the Kuroshio the Pco_2 is lowest (nearly 300 ppm). In the cold water region of the Yellow Sea, the Pco_2 is highest (nearly 370 ppm). In the south of about 30°N the Pco_2 does not exceed 320 ppm which is less than the pco_2. In the coastal area west of Kyūshū it is about 340 ppm. In the north of 30°N the Pco_2 is higher than the pco_2.

Longitudinal variations of the Pco_2 along 31°50′N are referred to changes of water temperature and pH *in situ* in Fig. 4. The highest Pco_2 occurs at the location of 125°42′E where the lowest water temperature is observed. This may reflect the properties of cold water extending southward from the Yellow Sea.

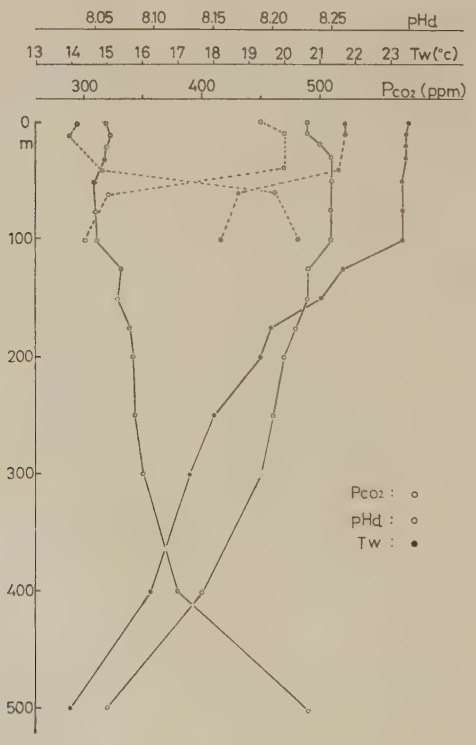

Fig. 5. Vertical distributions of the partial pressure of carbon dioxide in sea water (Pco_2) and water temperature and pH *in situ* at locations of 28°50′N, 128°30′E (solid line), and 29°43′N, 127°08′E (broken line).

As shown in the vertical profile of the water temperature, the warm water flowing northward branched from the Kuroshio (Tsushima Warm Current) is located at about 127°E, where the relatively low Pco_2, high water temperature and high pH *in situ* are observed. Variations of the Pco_2 and the water temperature do not correlate positively in the East China Sea because of the warm and the cold water masses originated from the different sources.

The vertical variations of the Pco_2 were examined at two representative points located at 28°50′N, 128°30′E, just south of the Kuroshio and at 29°43′N, 127°08′E, on the continental shelf. The vertical gradients of the Pco_2 in the ocean should be referred to in considerations of transfer through the air-sea interface and circulation of carbon dioxide in the ocean.

At the former location, the Pco_2 shows almost constant value to the depth of 100 m with a little fluctuation, and below the depth it increases gradually to the value of 500 ppm at the depth of 500 m. At the latter location, the Pco_2 increases sharply just below the thermocline and reached to the value of 490 ppm at the bottom of 100 m depth. The vertical variations of the water temperature and pH *in situ* corresponds well to the Pco_2 variations.

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The concentration of calcium and the calcium chlorosity ratio in the western North Pacific Ocean

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Abstract: The calcium concentration in sea water have been determined in the western North Pacific Ocean from January to March in 1967 and 1968. The observed area included the Kuroshio, the Subtropics, the North Equatorial Current, and the Equatorial Countercurrent. Horizontal distributions in the surface waters in 1968 and vertical distributions along the 137°E meridian in 1967 are described. The ratios of calcium to chlorosity have been determined. The ratios gradually increase in deep water under the salinity minimum layer except for the region of the Equatorial Countercurrent.

1. Introduction

It is well known that the chemical composition is almost constant throughout the world oceans. However, with the development of the analytical technique it has been evident that the composition of sea water has a slight variation in ionic ratios with depths and geographical positions (Miyake, 1939; Culkin and Cox, 1966).

With respect to the variation of calcium content in sea water, there are works by Sugawara and Kanamori (Kawasaki) (1958), Cox, Culkin, Greenhalgh and Riley (1962), and Tsunogai, Nishimura and Nakaya (1968).

In the present study, the present author intends to observe the distribution of calcium concentration of the western North Pacific Ocean and to clarify the relation between the distribution of calcium and the circulation of waters.

2. Samples and method of analysis

The sea water amples were collected from the western North Pacific Ocean on the cruises of R. V. *Ryofu Maru* for the CSK (Masuzawa, 1967a, 1968). The salinity was determined again at the time of analyses of calcium to ascertain the salinity change because the samples were stored in polyethylene bottles for several months, and the calcium concentrations were corrected.

Determinations of calcium were carried out by the titration method of TSUNOGAI *et al.* (1968) with a piston burette (manufactured by the Metrohm Co., Switz.) and a pipette which was designed in order to pipette a given volume of water sample. EGTA solution was standardized against a calcium carbonate of special reagent quality, low in alkalies (Kanto Chemical Co., Inc.). Analytical error is considered to be less than 0.2%.

3. Results and discussions

3-1. Horizontal distribution in the surface waters

Calcium concentrations in the surface waters in the observed area are within 398-418 mg/*l*. The higher values are observed in the area south of the Kuroshio (418 mg/*l*) and the region along 20°N (414 mg/*l*). Both areas seem to coincide with the northern and southern edges of the Subtropic region. Other higher values are found in the waters of the Equatorial Countercurrent (415-417 mg/*l*). The lowest value is found in the water east of Mindanao Island (398 mg/*l*). The Subtropic waters show a low value of 410 mg/*l*. (Fig. 1 a)

Calcium chlorosity ratios vary between 2063×10^{-5} and 2125×10^{-5}. The Sub-

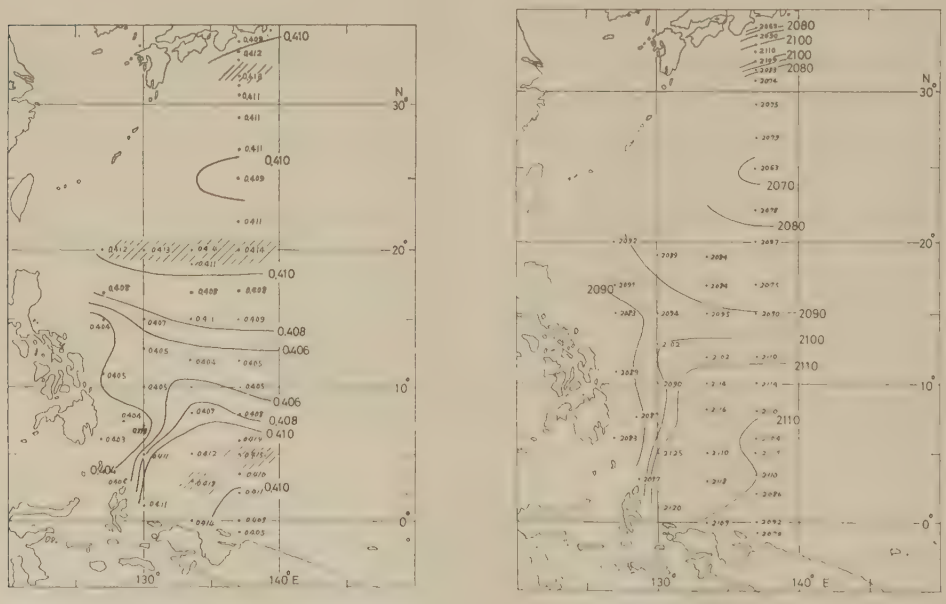

Fig. 1. Distribution of calcium concentration in g/*l* (a), and calcium chlorosity ratio in $Ca/Cl\times10^5$ (b) in the surface water from January to March 1968.

tropic waters show the lower values than 2080×10^{-5} in connection with high salinity. It may be conceivable that the calcium is removed from water by living organisms in the stagnant waters. In the region of the North Equatorial Current the ratios increase gradually toward the south from 2080×10^{-5} to 2100×10^{-5}. The higher values than 2110×10^{-5} are found in the region of the Equatorial Countercurrent. This may be due to addition of calcium from the land or the South Pacific. (Fig. 1 b)

3-2. Vertical distribution along the 137°E meridian

The vertical variations of calcium concentration and of calcium chlorosity ratio along the 137°E meridian in January 1967 are shown in Fig. 2. The maximum deviation of the calcium concentration in this section does not exceed 2%, which is ten times of the analytical error. The layer with the minimum calcium content extends toward south to 15°N. This layer becomes gradually shallow southward from 30°N to 15°N.

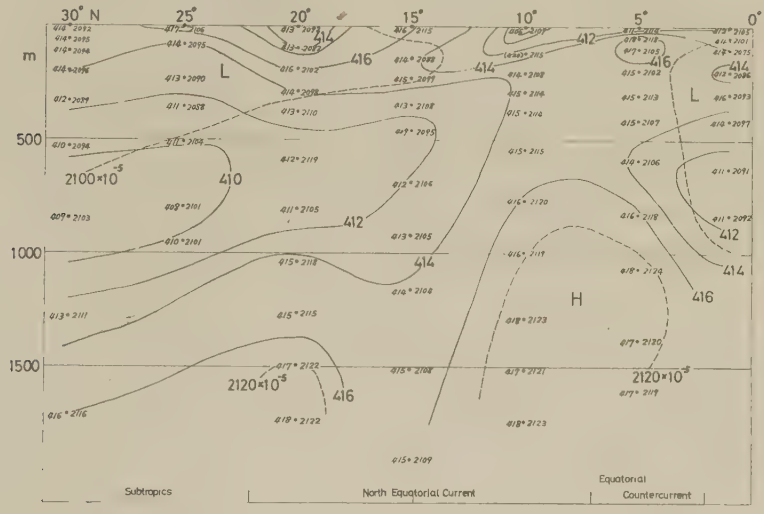

Fig. 2. Distribution of calcium concentration (mg/l) and calcium chlorosity ratio ($Ca/Cl\times10^5$) in a vertical section along the 137°E meridian in January 1967.

Masuzawa (1967 b) pointed out that the North Equatorial Current can be divided into two parts. Similarly to his result, near the location of 15°N the North Equatorial Current has a different calcium concentration between nouthern and southern parts. Calcium contents are rich in the region of 10°N to 5°N at

the depth deeper than 800 m. Near the equatorial region a low values (412 mg/l) are observed.

On the calcium chlorosity ratios waters near the equatorial region have also low values and next to the region the Subtropic one have low values. A similar pattern of the alkalinity in the same section has been reported (AKIYAMA *et al.*, 1968). The low calcium content may be caused by the high productivity in the sea water. Such sea water show a low specific alkalinity which is caused by the supersaturation in the carbonate content.

When the titration alkalinity chlorosity ratio changes corresponding to the calcium chlorosity ratio by the solution or precipitation of calcium carbonate, the calcium concentration in sea water can be calculated by the equation of WATTENBERG (1936);

$A=2\mathrm{Ca}-0.93\times\mathrm{Chlorosity}$.

Here, A represents the titration alkalinity in m. equivalents/l; Ca represents the calcium concentration in mg•atoms/l. The relation of the calculated values to the measured ones is shown in Fig. 3. In the diagrm, the points are scattered

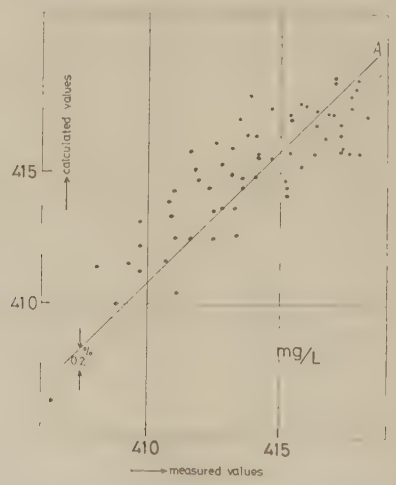

Fig. 3. The relationship between the measured value and the calculated value in the concentration of calcium along the 137°E meridian in January 1967.

around line A and give a maximum deviation of only 1%. The calculated values are higher by about 0.2%. This difference, however, do not exceed the analytical errors.

Culkin and Cox (1966) confirmed that the mean calcium chlorinity ratio for all the deep samples was higher than for the surface samples in some oceans. Tsunogai *et al.* (1968) reported the similar results. Fig. 4 shows the relation of

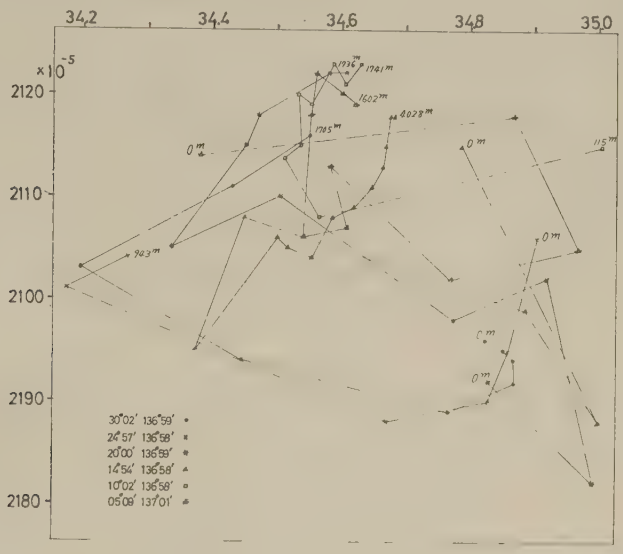

Fig. 4. The relationship between the calcium chlorosity ratio and salinity.

the calcium chlorosity ratio to the salinity. Although the values are scattered the ratios show a tendency to increase in deep water below the salinity minimum layer where the solubility of calcium carbonate increases with lowering temperature and increasing pressure.

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Note on a species of *Calanoides* (Copepoda, Calanoida) from the western North Pacific

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and

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(Received July 25, 1969)

Abstract: Plankton collections were made in the western North Pacific from January to March 1967 on board the R. V. *Ryofu Maru* for the CSK. From these samples a new species of Calanoid Copepoda, *Calanoides philippinensis* n. sp., was described. This species closely resembles *Calanoides carinatus* (KRÖYER), but could be distinguished from it by the shape of genital opening in female and the structure of the fifth pair of legs.

A large number of *C. philippinensis* occurred in the region where colder deep water ascended to upper layers. This species seems to live chiefly in the South China Sea and to be dispersed northward by the midwater of the Kuroshio through the Bashi Channel.

1. Introduction

Five species of *Calanoides* have been reported by many authors from various localities. They are *C. carinatus* KRÖYER (=*brevicornis* GIESBRECHT), *C. patagoniensis* BRADY, *C. natalensis* BRADY, *C. acutus* GIESBRECHT and *C. macrocarinatus* BRODSKY. Among them *C. acutus*, the Antarctic species, has been described in details by VERVOORT (1951). *C. carinatus* is very widely distributed in the Atlantic and Indo-Pacific Oceans, and has been recorded by GIESBRECHT (1892), FRÜCHTL (1923), FARRAN (1929), VERVOORT (1947, 1951) and TANAKA (1937, 1960, 1964). According to SEWELL (1947), the immature specimen described by TANAKA from Suruga Bay is the synonym of *C. patagoniensis*. On the other hand *C. patagoniensis* has rather limited localities in distribution and has been recorded from the west coast of South America by BRADY (1883) and GIESBRECHT (1892), from the tropical Atlantic off Rio de Janeiro by FARRAN (1929) and from the Gulf of Oman and Arabian Sea by SEWELL (1947). These two species resemble each other

not only in general appearance and in the structure of oral appendages, but also in habitat. *C. natalensis* is very imperfectly known. As for *C. macrocarinatus* the detailed study is in press (BRODSKY's personal communication).

Recently, the specimens of *Calanoides* were found in the material collected on board the R. V. *Ryofu Maru* of the Japan Meteorological Agency, during her research cruise to the western North Pacific from January to March, 1967 (CSK). The female specimens are very closely allied to *C. carinatus*. On the other hand, the male specimens come near to *C. patagoniensis* in the structure of the 5th pair of legs.

Plankton samples were collected with two types of net at stations shown in Fig. 4. Norpac net was hauled vertically from about 150 m depth to the surface, and the vertical haulings of the automatic opening- and -closing multiple net from 1,000 m depth to the surface with intervals of 250 meters were made. Counting of specimen was made on 4th stage copepodite, 5th stage copepodite and adults. The numbers of male and female specimens collected were 16 inds. and 64 inds., respectively.

2. Characters of female and male

2-1. Female (Length 2.55—2.79 mm)

The cephalothorax and abdomen in the proportional lengths are as 81 to 19. The frontal margin of the head is triangular in shape, and has a slight chitinous thickening on the anterior end when viewed from the dorsal. In lateral aspect it is rather broadly rounded. The lateral margins of the last thoracic segment are narrowly rounded and slightly pointed at the distal end. The rostral filaments are slender.

The abdominal segments and furca are in the following proportional lengths:

Segment	1–2	3	4	5	Furca
	42	15	13	11	19=100

The genital segment is nearly as long as width and swollen laterally at the middle section. The genital area protrudes below. The genital opening resembles that of *C. carinatus* but can be distinguished from that of the latter by differences in structure to some extent. The furcal rami is 1.4 times as long as width.

The 1st antenna extends to the distal margin of the genital segment. It measures 2.49 mm in length, so the proportional lengths of the antenna to the

Fig. 1. *Calanodes philippinensis* n. sp., female.
a. head, lateral view (2.64 mm); b. head, dorsal view (2.74 mm); c. abdomen, lateral view (2.64 mm); d. abdomen, lateral view (2.74 mm); e. genital opening, ventral view (2.64 mm); f. genital opening, ventral view (2.84 mm); g. rostrum; h. 2nd antenna; i. mandible blade; j. 1st maxilla; k. 2nd maxilla; l. maxilliped; m. 1st leg; n. 2nd leg; o. 3rd leg; p. 4th leg; q. 5th leg.

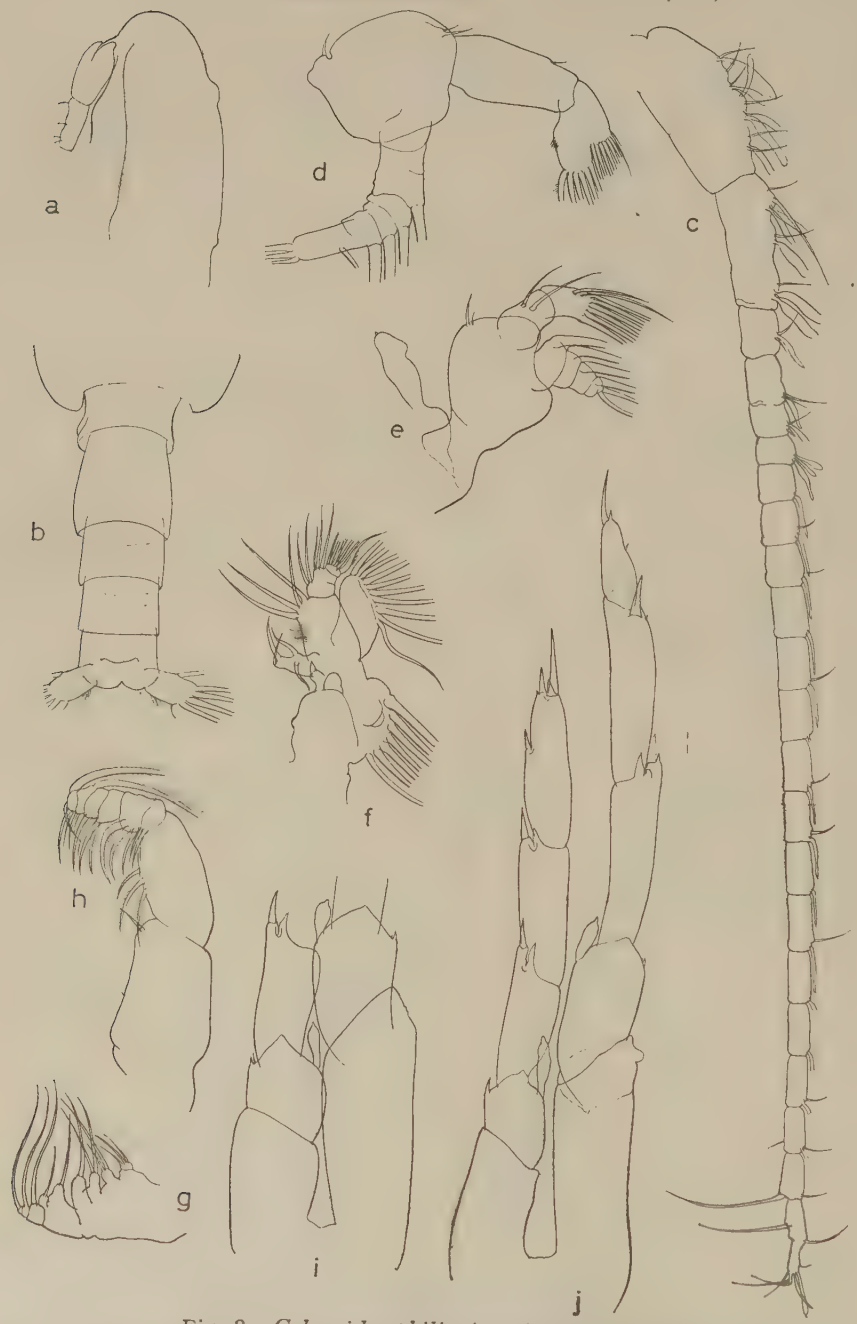


Fig. 2. *Calanoides philippinensis* n. sp., male.
a. head, lateral view (2.65 mm); b. abdomen, dorsal view (2.65 mm); c. 1st antenna (right); d. 2nd antenna; e. mandible; f. 1st maxilla; g. 2nd maxilla; h. maxilliped; i. 5th pair of legs (2.60 mm); j. 5th pair of legs (2.65 mm).

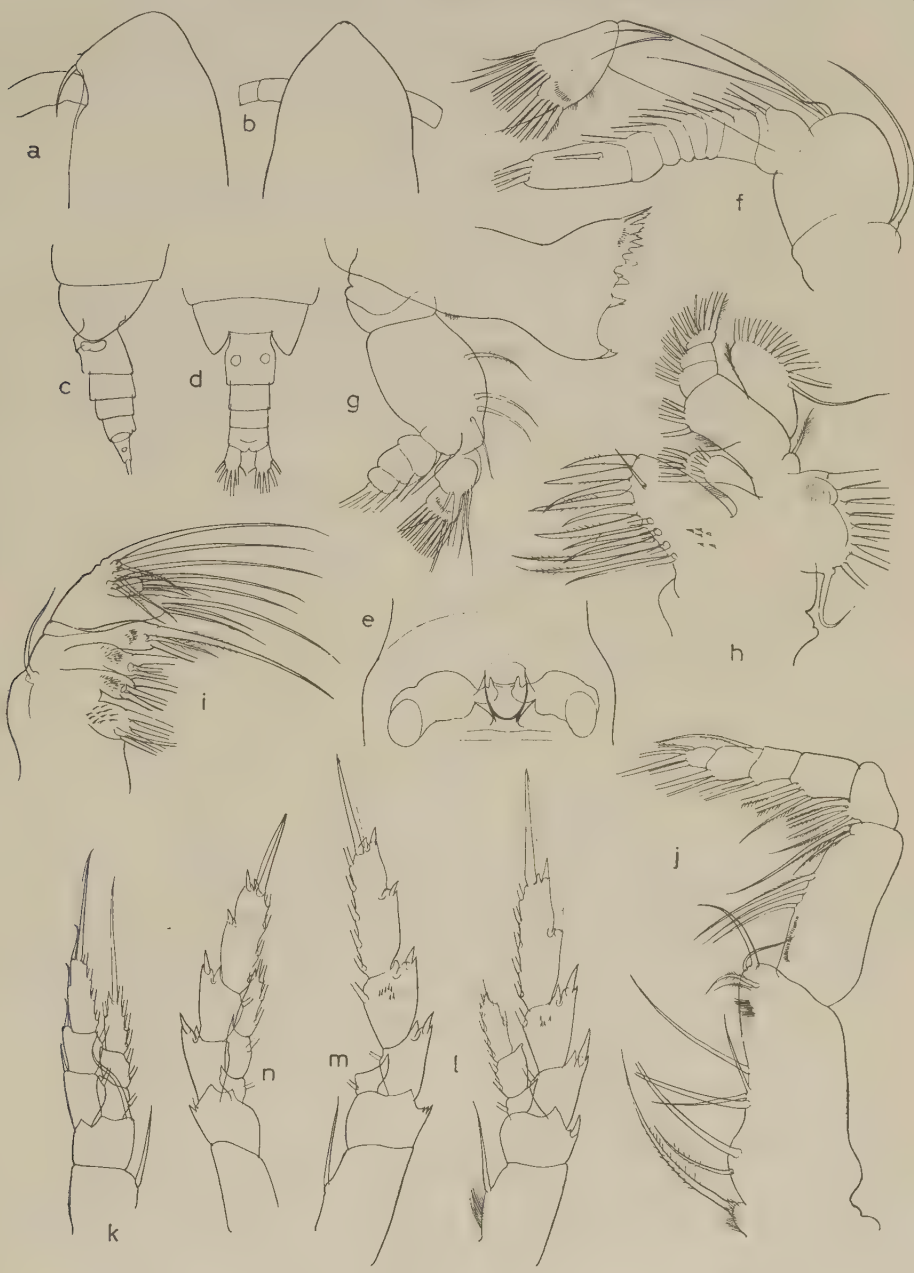


Fig. 3. *Calanoides carinatus*, female (4.20 mm) of Antarctic specimen.
a. head, lateral view; b. head, dorsal view; c. abdomen, lateral view; d. abdomen, dorsal view; e. genital opening, ventral view; f. 2nd antenna; g. mandible; h. 1st maxilla; i. 2nd maxilla; j. maxilliped; k. 1st leg; l. 2nd leg; m. 4th leg; n. 5th leg.

body are as 90 to 100. The segments are in the following proportional lengths:

Segment	1	2	3	4	5	6	7	8–9	10	11	12	13	14	15	16	17
	73	64	40	36	36	36	36	64	32	28	32	30	40	43	40	43
	18	19	20	21	22	23	24	25								
	43	40	36	44	40	36	32	40=1000								

The 2nd antenna, mandible, 1st and 2nd maxillae and maxilliped are as those of *C. carinatus* described by VERVOORT (1947).

The 1st to 5th swimming legs have each 3-jointed exopod and endopod. In the 2nd to 4th legs, the 2nd segment of the exopod is furnished with a group of spinules on the posterior surface. These spinules are also found in the specimens of *C. carinatus* taken from the Antarctic region. The proportional lengths of the outer margin of the 3rd segment of the exopod divided by the marginal spine in the specimens measuring 2.64 and 2.74 mm are as follows:

	2.64 mm		2.74 mm	
	proximal	distal	proximal	distal
2nd leg	20	20	20	19
3rd leg	27	20	25	20
4th leg	29	19	28	21

These figures do not coincide with those for *C. carinatus* given by GIESBRECHT and SCHMEIL (1898).

2-2. Male (Length 2.65 mm)

The cephalothorax and abdomen in the proportional lengths are as 77 to 23. In lateral aspect the frontal margin of the head is evenly rounded. There is a small swelling on the dorsal surface just opposite of the place of the attachment of the 2nd antenna. The last thoracic segment is rounded on the lateral margins.

The abdominal segments and furca are in the following proportional lengths:

Segment	1	2	3	4	5	Furca
	16	28	16	14	14	12=100

The furcal rami are 1.5 times as long as width.

The 1st antenna is 21-jointed and extends to the distal margin of the furca. It measures 2.47 mm in length, so the proportional lengths of the antenna to the body are as 94 to 100. The segments are in the following proportional length:

Segment	1–2	3–4–5	6	7	8	9	10	11	12	13	14	15	16	17
Left	138	97	44	41	32	24	32	28	32	44	41	41	44	41
	18	19	20	21	22	23	24–25							
	45	41	41	45	41	41	67=1000							

The proximal segments are furnished with long sensory filaments.

The 2nd antenna is slightly deformed. The exopod and endopod are of nearly equal lengths. The exopod is 6-jointed and is furnished with 4 marginal setae.

The mandible degenerates much. The cutting edge is provided with a single tooth.

The 1st maxilla degenerates. The 1st inner lobe has no seta; the 2nd lobe carries a small pointed lamellous process; the 3rd lobe has 4 short setae. The endopod is furnished with 3, 2 and 3 setae on the segments 1 to 3. The 2nd basal has 4 setae. The outer lobe has 7 setae. The exopod has 10 setae. The 2nd maxilla degenerates in the armature of setae on the lobes. The maxilliped is robust.

The 1st leg has the similar structure to that of the female. The 2nd to 4th legs are broken off in the present specimen.

The 5th pair of legs has 3-jointed exopod and degenerated endopod on each side. In the left leg the 1st basal segment is long, and extends as far as the distal margin of the 2nd basal segment of the right leg; the 2nd basal segment reaches to the distal margin of the 1st segment of the exopod of the right leg, and is furnished with a rudimentary endopod just as is observed in the same leg of *C. patagoniensis;* the 1st and 2nd segments of the exopod are of nearly equal lengths; the 3rd segment of the exopod is half as long as the 2nd, and is furnished with a terminal spine which is about half the length of the 3rd segment of the expopod. The right leg is provided with a rudimentary endopod which degenerates in structure compared with that of *C. patagoniensis.* The endopod is incompletely 3-jointed, and reaches half way along the 1st segment of the exopod of the right leg; these three segments are of nearly equal lengths. and carry no setae. The terminal spine of the exopod is about half the length of the 3rd segment of the exopod of the same leg.

2-3. Remarks

As the present specimen is very closely related to *C. carinatus,* so it is with some hesitation to separate the present female specimens from *C. carinatus.* These species resemble each other not only in general appearance but also in the structure of the mouth parts and swimming legs. In general, one of the most important character in the systematic study of copepods is the structure of the 5th pair of legs. It is unfortunate that in *Calanoides* the female 5th pair

of legs does not play an important role in the identification of the species. The specific characters have been mainly focussed on the shape of the forehead and the lateral margins of the last thoracic segment, and the proportions of the outer margin of the 3rd segment of the exopod divided by the marginal spine in the 2nd to 4th legs. However, these characters are not useful, as the similar features are met with both in *C. carinatus* and in the present specimen. The other character separating the species is the structure of the genital opening. The genital opening of the present specimen has been different in structure to some extent from that of *C. carinatus*. It differs also from that of *C. patagoniensis* described by BRODSKY (1967). On the other hand, the male has remakable character in the structure of the mouth parts and the 5th pair of legs. According to VERVOORT (1951) the male of *C. acutus* has the degenerated mouth parts not only in size but also in the number of setae on the various lobes of the 1st and 2nd maxillae. In the male of the present specimen the oral parts degenerate compared with those of *C. acutus*, especially in the structure of the mandible blade, 1st to 3rd inner lobes of the 1st maxilla and in the armature of the lobes of the 2nd maxilla. The 5th pair of legs of *C. acutus* approaches in the structur of the endopod to that of *Calanus*. The endopod degenerates in *C. carinatus* and further in *C. patagoniensis* which is shown by the absence of setae on the 1st and 2nd segments of the endopod of the right leg. The process of degeneration most goes in the present specimen, represented by the complete absence of setae on the endopod of the same leg. These characters of the male in the present specimen will prove that the specimen is quite differrent from other species mentioned above. Therefore, authors give the name, *Calanoides philippinensis*, to the specimen studied here.

3. Distribution of *Calanoides philippinensis*

3-1. The section along 137°E

The divided vertical hauls from 1,000 m to the surface were made at four stations shown in Fig. 4. At Ry 3037 (30°02′N) 5th stage copepodites were collected in the layer of 750 to 1,000 m in small numbers, being 25 inds. per 1,000 cubic meter of sea water. At Ry 3052 (14°54′N), Ry 3057 (10°02′N) and Ry 3062 (5°09′N) no specimens were collected. Ry 3037 is located in the Kuroshio Countercurrent, Ry 3052 and Ry 3057 are in the North Equatorial Current and Ry 3062 in the Equatorial Countercurrent (MASUZAWA, 1967).

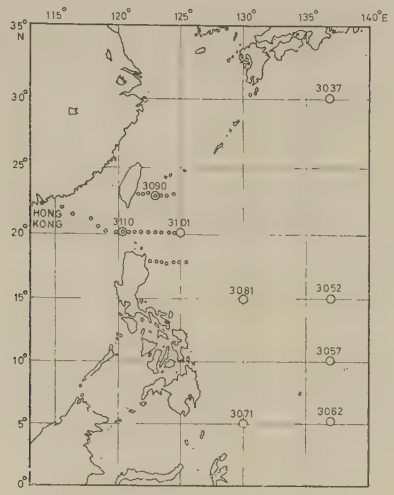


Fig. 4. Locations where plankton collections were made. Large circles indicate stations where the divided vertical hauls were made from 1,000 m depth to the surface with intervals of 250 meters with the automatic closing- and -opening multiple net and small circles indicate stations where vertical hauls were made from about 150 m depth to the surface with Norpac net.

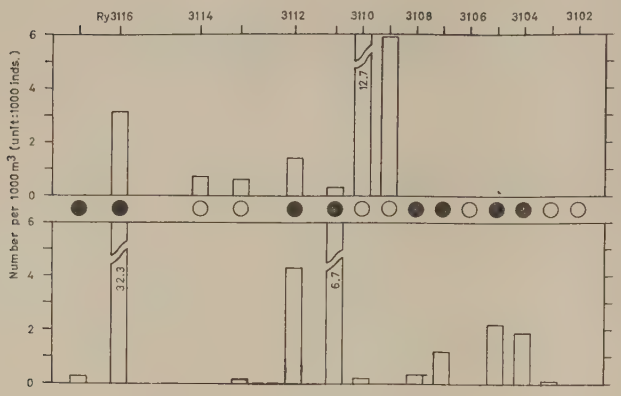
3-2. The section along 130°E

The divided vertical hauls as in the section along 137°E were made at two stations, Ry 3071 (5°04′N) and Ry 3081 (14°59′N), but no specimens were taken from any layer (Fig. 4). From the planktological point of view, it has been confirmed by KAWARADA *et al.* (1968) that these stations were located in the Equatorial Countercurrent and the North Equatorial Current, respectively.

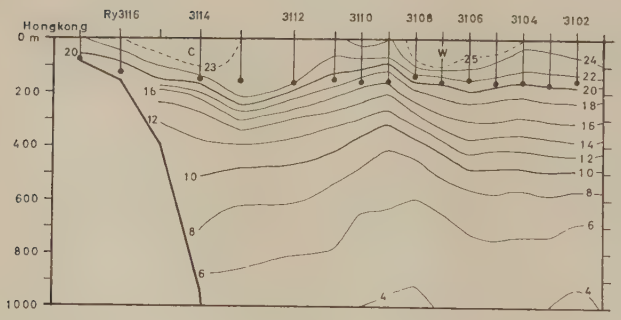
3-3. The section from Ry 3101 (20°10′N, 124°57′E) to Hongkong

At sixteen stations vertical hauls from a maximum of 170 m depth to the surface were made with Norpac net. The distributional feature of the species is demonstrated in Fig. 5. From the figure it is evident that *Calanoides philippinensis* is restricted in the South China Sea. The largest amount appears at Ry 3110 (20°10′N, 119°58′E), and fairly large amount at Ry 3109 (20°12′N, 120°58′E). At Ry 3110 the population consists of both 5th and 4th stages copepodites, the former exceeding the latter in proportion. Adults occur at Ry 3109, Ry 3112 (20°19′N, 119°01′E) and Ry 3116 (21°36′N, 116°17′E) in small numbers.

In this section, the divided vertical hauls from 1,000 m to the surface were


Ry3116
3114
3112
3110
3108
3106
3104
3102
Number per 1000m³ (unit:1000 inds.)
12.7
32.3
6.7

a


Hongkong
Ry3116
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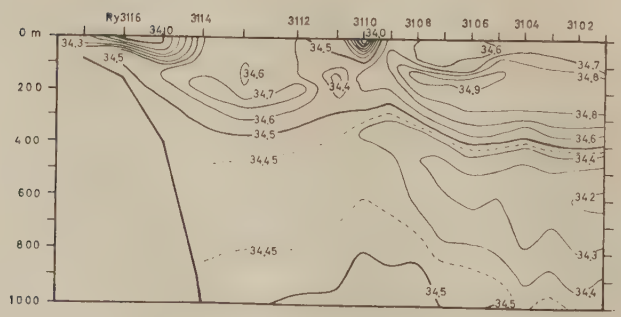
Ry3116
3114
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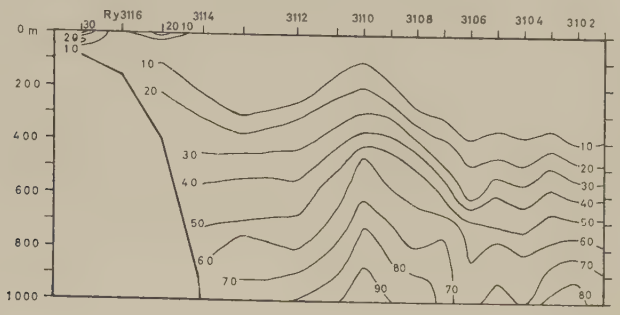
1000

c



d

Fig. 5. Distributions of *Calanoides philippinensis* (upper) and *Pleuromamma* (lower) (a) and vertical sections of temperature (b), salinity (c) and silicate-Si (μg-atmos/l) (d) in the section from Ry 3101 to Hongkong. White circles indicate daytime collections, black circles night collections and small black circles the depth of Norpac net lowerd.

made at Ry 3101 and Ry 3110. At the former station only 5th stage copepodites occur in deeper two layers of 500 to 750 m and of 750 to 1,000 m, but the numbers of individuals are extremely small, being 75 and 125 inds., respectively. At the latter station the greatest amount was collected. As shown in Fig. 6, the

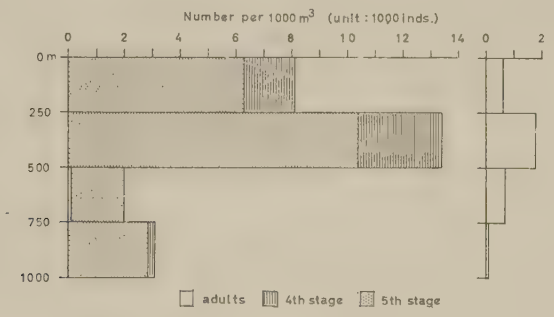


Fig. 6. Vertical distributions of *Calanoides philippinensis* (left) and *Pleuromamma* (right) at Ry 3110. Collection was made in the daytime on February 23, 1967.

population is concentrated upper two layers (0-250 m, 250-500 m) with the highest concentration in the layer of 250 to 500 m, where the stage composition resembles each other, 5th stage copepodites exceeding 4th stage copepodites in proportion.

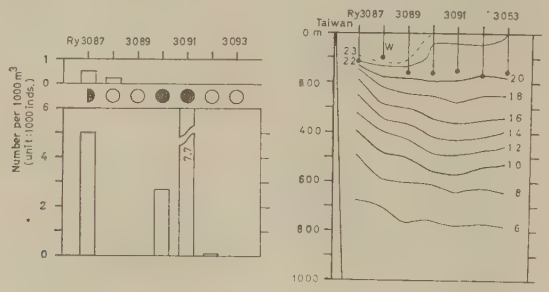
In respect to the diurnal vertical migration, Fig. 5 suggests that *C. philippinensis* does not present a clear pattern of the diurnal movements in contrast

with the behavior of *Pleuromamma*, which has been well known as the species presenting the conspicuous diurnal vertical migration in the warm oceanic waters (FURUHASHI, 1961).

It is indicated in the vertical distributions of water temperature, salinity and silicate that the deeper water ascends to the upper layers at Ry 3109 and Ry 3110, being the western margin of the Kuroshio as shown in Fig. 5. It is interesting fact that the region concentrated with *C. philippinensis* coincides with ascending water region as mentioned above.

3-4. The section east of Taiwan

In the vertical hauls from a maximum of 170 m depth to the surface, a small number of 5th stage copepodites were collected at Ry 3087 (23°01′N, 121°34′E) and Ry 3088 (23°00′N, 122°02′E) located at the margin of the Kuroshio, where colder deep water ascends to upper layers (Fig. 7). The occurrence of 5th copepodites seems to have no connection with the diurnal vertical migration as in the case of the section from Ry 3101 to Hongkong.



a b

Fig. 7. Distributions of *Calanoides philippinensis* (upper) and *Pleuromamma* (lower) (a) and vertical section of temperature (b) in the section east of Taiwan. White circles indicate the daytime collections, black circles the night collections and black semicircle twilight collection.

At Ry 3090 (22′58′N, 123°02′E) the divided vertical haul was made, and extremely small number of 5th stage copepodites, 25 inds. were taken from the layer of 750 to 1,000 m.

3-5. The section east of Luzon Island

Vertical hauls from a maximum of 220 m depth to the surface were made. No specimen were collected even at stations close to Luzon Island, although

the vertical profile of water temperature in this section is similar to that in the section east of Taiwan.

3-6. *Discussion*

From the pattern of the vertical distribution of *Calanoides philippinensis*, it might be thought that this species can be grouped in meso-palnktonic species. In general, meso-planktonic species is accepted as to present the diurnal vertical migration, but it does not occur in 5th stage copepodites of this species, so that the occurrence of 5th stage copepodites in the various layers may have no concern with the diurnal vertical migration. In respect to water temperature which is one of main elements to restrict the distribution of zooplankton, 5th stage copepodites have wide torelance, as they are collected both in the deep water with a temperature of 4 to 7°C and in the shallow water with a temperature above 22°C.

It thus appears that the occurrence of the large amount is due to the phytoplankton feeding associated with the hydrographic condition. In the South China Sea, the stations where great amount of 5th stage copepodites occurred are located in the region where the ascending of water from the deep takes place, so that it is expected that heavy phytoplankton production takes place with the supply of rich nutrients. They contained light-brown globules of oil. This fact may suggest that they has taken a large amount of phytoplankton. In this region large amount of diatoms are present (Mr. A. SANO's personal communication). A similar phenomenon has been observed on *C. carinatus* at the west side of Africa by BAINBRIDGE (1960). He reported that the 5th stage copepodites occurs plentifully in the surface water along the edge of the shelf off Ghana, where the surface temperature is lower in comparison with that of neighbouring waters, and they contain orange globules of oil. Thus *C. philippinensis* appears to spend parts of life cycle at shallow water until early young copepodites develop to 5th stage copepodites, and they probably descend into deep water.

C. philippinensis does not seem to be living in the North Equatorial Current and in Equatorial Countercurrent. This animal mainly lives in the intermediate depth in the South China Sea, and a part of population may be carried northward through the Bashi Channel by the midwater of the Kuroshio, so far as the present survey is concerned.

One species of *Calanoides*, which is probably identical with *C. philippinensis* is collected from the surface water at the edge of the continental shelf in the

East China Sea in winter (Mr. M. MATSUZAKI's personal communication). In this region phytoplankton stock is larger in winter than in summer (KAWARADA *et al.*, 1966, 1968), so that it is considered that the occurrence of *Calanoides* in the surface is related to phytoplankton feeding.

Acknowledgment: We would like to thank Dr. Y. KAWARADA of Japan Meteorological Agency for a critical reading of manuscript.

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The preliminary report of the third cruise for CSK, Jauuary to March 1969

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and

Tsutomu Sawara

Japan Meteorological Agency

(Received July 5, 1969)

The third cruise for the Cooperative Study of the Kuroshio and Adjacent Regions (CSK) was made in January to March 1969 east of the Philippines by the R. V. *Ryofu Maru* of the Japan Meteorological Agency (JMA). The survey consists of two long meridional sections along 137°E and 130°E, two short meridional sections along 133°30′E, and 127°E, and six short sections east of the Philippines (Fig. 1). As the sea state was more quiet this year than the last two years, the prearranged plan was almost completed.

1. Cruise itinerary

Port	Arrival	Departure
Tokyo, Japan	—	January 14
Cebu, Philippines	February 7	February 11
Naha, Okinawa	March 1	March 5
Keelung, Taiwan	March 15	March 20
Tokyo, Japan	March 24	—

2. Personnel

Captain: K. Shirakawa
Officers and crew: 35 persons
Scientists: Chief, H. Akamatsu, Marine Division, JMA
Phys. Oceanography: T. Sawara, Marine Division, JMA
I. Fujiwara, Marine Division, JMA

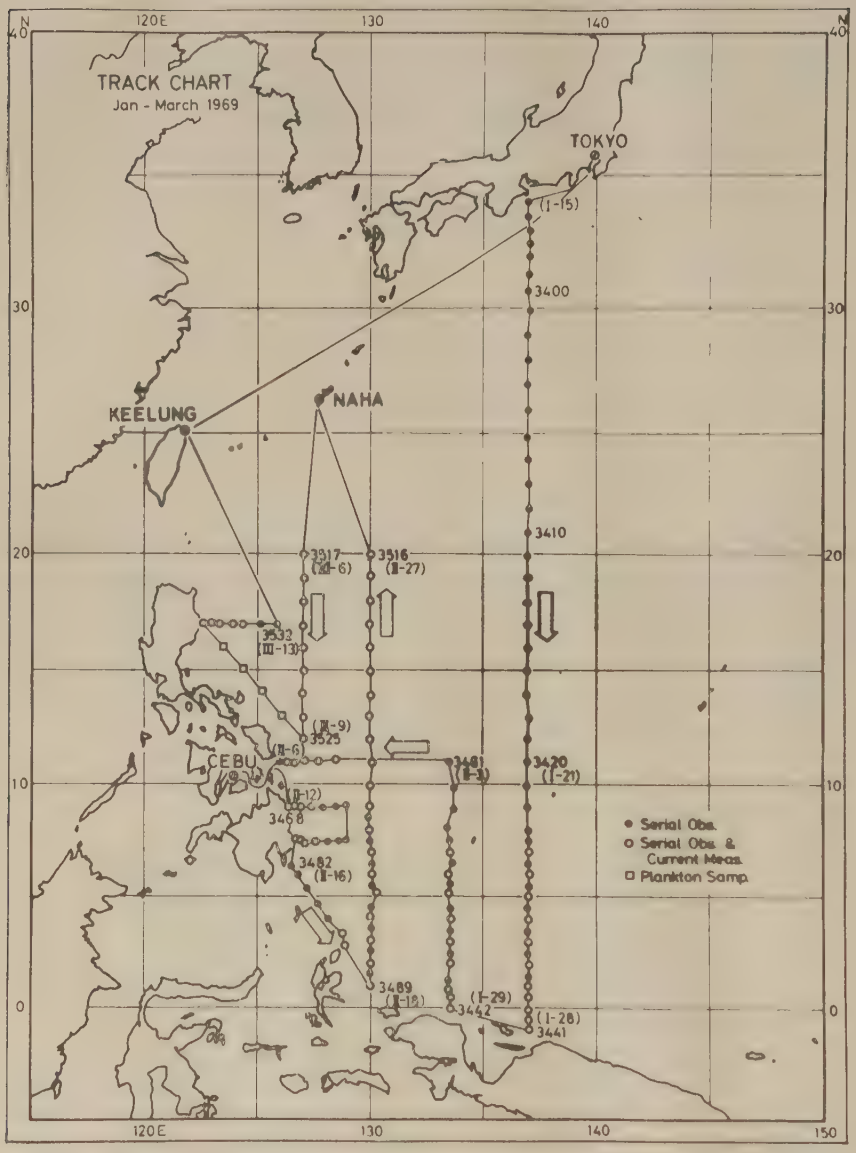


Fig. 1. Track chart of the third cruise for CSK made by *Ryofu Maru* of the JMA in January to March 1969. The dot, open circle and square indicate the serial observation only, serial observation and current measurement, and plankton sampling respectively.

T. Urashima, Marine Division, JMA
N. Sato, Marine Division, JMA
S. Hosoda, Hakodate Marine Observatory
T. Yasuoka, Maizuru Marine Observatory
T. Tsuchida, Nagasaki Marine Observatory

Chemistry : T. Akiyama, Marine Division, JMA
T. Sagi, Marine Division, JMA
T. Yura, Marine Division, JMA
N. Baba, Kobe Marine Observatory

Planktology : Y. Kawarada, Marine Division, JMA
K. Furuhashi, Marine Division, JMA
A. Sano, Marine Division, JMA

Meteorology: S. Hayashi, Marine Division, JMA
H. Takano, Marine Division, JMA
A. Yamashita, Marine Division, JMA

UNESCO shipboard fellow:
Choo Kyo Sung, Hydrographic Office, Korea
Claro S. Doctor, Weather Bureau of Manila, Philippines
Hiran Sartrasook, Naval Weather Forcasting Station, Thailand

3. Type of observation

Number of stations: 139

One hundred and two stations for hydrographic cast deeper than 1,000 m were made at intervals of 60 miles on the prearranged lines. In addition, 37 shallow stations (less than 500 m) were taken at the middle between adjacent deep stations south of 8°N and near the Philippines.

Sampling depth:

North of 20°N: 50, 100, 150, 200, 300, 400, 500, 600, 700, 800, 900, 1,000, 1,100 (1,500, 2,000, 2,500, 3,000 m).

8°N to 20°N : 25, 50, 75, 100, 125, 150, 175, 200, 250, 300, 350, 400, 500, 600, 700, 800, 900, 1,000, 1,100 (1,500, 2,000, 2,500, 3,000m).

South of 8°N and near the Phillippines:
25, 50, 75, 100, 125, 150, 175, 200, 225, 250, 275, 300, 350, 400, 500, 600, 800, 1,000, 1,200, (1,500, 2,000, 2,500, 3,000 m).

Max. sampling depth: 4,939 m.

BT observation: BT lowerings were made at each hydrographic station and near the Philippines (143 station).

Current measurement: Current measurements were made with GEK and current meters.

11 stations between 20°N in 137°E: With GEK.

69 stations in Fig. 1: In the upper 500 m with the TS-II type self-recording-currentmeter by means of two currentmeter method.

Bottom topography: Soundings with the echo sounder at each hydrographic station.

Chemical element: Chemical elements determined on board are salinity (139 sts.), oxygen (139 sts.), phosphate-P (102 sts.), silicate-Si (102 sts.), nitrite-N (46 sts.), nitrate-N (55 sts.), pH and total alkalinity (6 sts.), partial pressure of carbon dioxide (Pco_2) in upper 300 m (20 sts.), and Pco_2 in surface water (139 sts.). Water was sampled from the surface and sub-surface depth for the determination of total-P (18 sts.), calcium (18 sts.), trace metal (18 sts.) and gross β radio activity (3 sts.).

Biological observation: The Norpac Net (bolting cloth GG54, 0.33 mm mesh) was hauled vertically from the estimated depth of 150 m at 143 stations for zooplankton standard sampling. The vertical haulings with the automatic opening-and-closing multiple net were also made from those four layers (0-250 m, 250-500 m, 500-750 m, 750-1,000 m) at 26 stations. One-liter surface water was sampled at 139 stations for the quantitative determination of microplankton. The sample water was also drawn from Nansen bottles at 12 depths in the upper 250 m at 13 stations. Chlorophyll and phaeophytin in the upper 250 m were measured with a fluorophotometer (HITACHI FPL-2) at 93 stations.

Meteorological observation: Routine maritime meteorological observations were carried out 8 times a day, and the measurements of cloud and rainfall area with the high-power weather radar, and of radiation from sky and sea, and the upper air ob-

servations with radio sonde and pilot balloon were sometimes made in the equatorial region. And also the pco_2 in the atmosphere was recorded underway.

4. Results

The general features of current patterns based on the direct measurement are as follows;

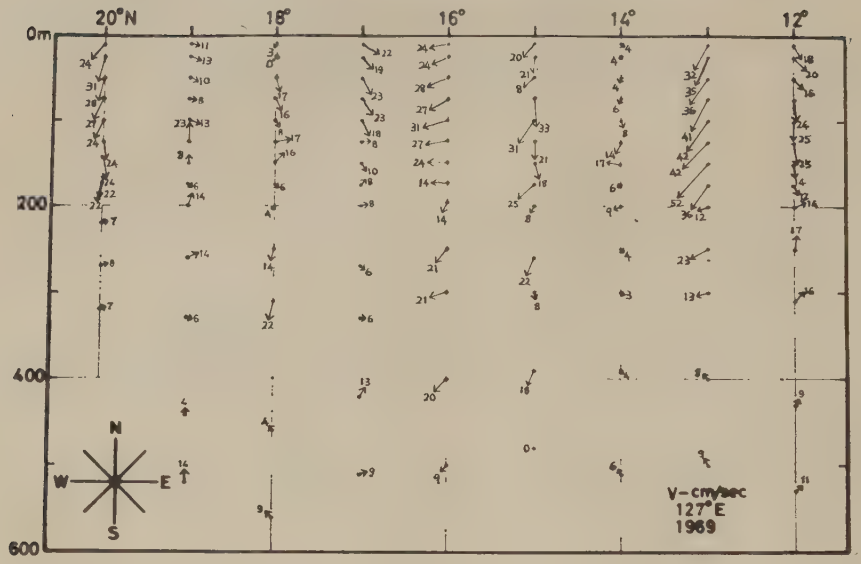
(1) At 130°E, the North Equatorial Current flows westward between 7°30′N and 18°30′N with about 1 kt at the sea surface, and the center of the current is found at 13°N, and the westward current extends to the depth of about 500 m. The strong currents are found at 10°N, 13°N and 16°N both in 130°E and 127°E (Fig. 2 a, b).

(2) The Equatorial Countercurrent flows eastward between 7°30′N and 3°30′N, and the maximum speed, 1.7 kt, is found at 5°N in 133°30′E and 137°E (Fig. 2 c, d).

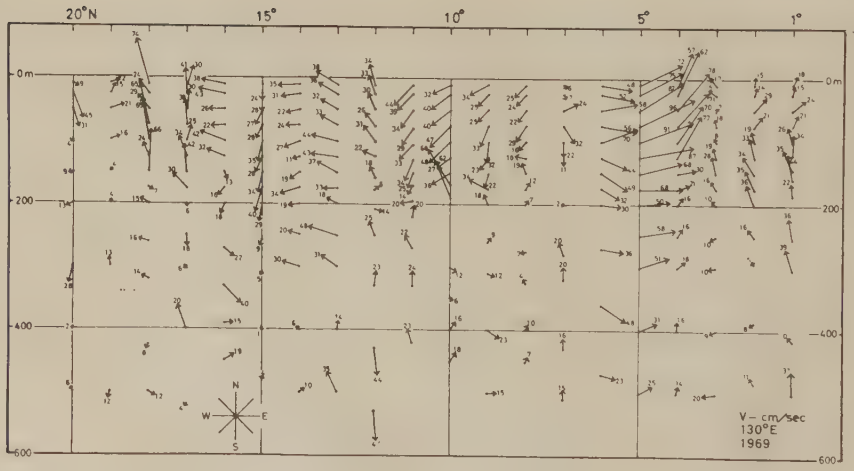
The eastward current in depths from 200 to 300 m at 3°N and the equator at the 137°E section seems to be the beginning of the Equatorial Undercurrent. Similar eastward currents near the equator were observed in 1967 and 1968, and the speed is about 1.0 kt in 1969 like in 1968, and about 3 kt in 1967 (Masuzawa, 1968).

(3) The westward South Equatorial Current as high as 1.5 kt was found in the 75-300 m layer between 1°S and 0°30′S, north of New Guinea. The South Equatorial Current at 137°E is lower in speed and narrow in 1969, though the current is covered by the shallow eastward flow of the New Guinea Coastal Current in 1969 like in 1968. The South Equatorial Current carries the high saline Tropical Water from the South Pacific to the north sea of New Guinea (Fig. 3). The Tropical Water from the South Pacific near the equator at 137°E is lower in salinity and smaller in volume of the saline water exceeding 35.5‰ in 1969 than in 1968. The low-salinity waters are found in depth of 200 m at 2°N in 133°30′E, and 180 m at 0°30′N in 137°E (Fig. 3). These waters may be of the North Pacific origin. A similar small watermass was observed in the 133°30′E, and 137°E sections in 1968 (Masuzawa, 1968).

(4) The Mindanao Current flowed nearly due south in the narrow coastal zone with the width of about 30 miles. The maximum speed measured by the current meter is 2.6 kt referred to the velocity at 700 m depth. The current measurement was done five times during 24 hours at 11°N, 127°20′E, near the

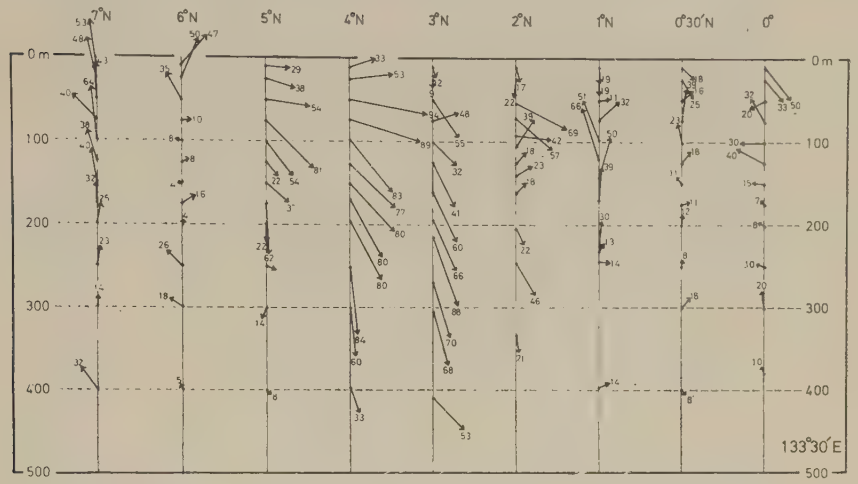


a

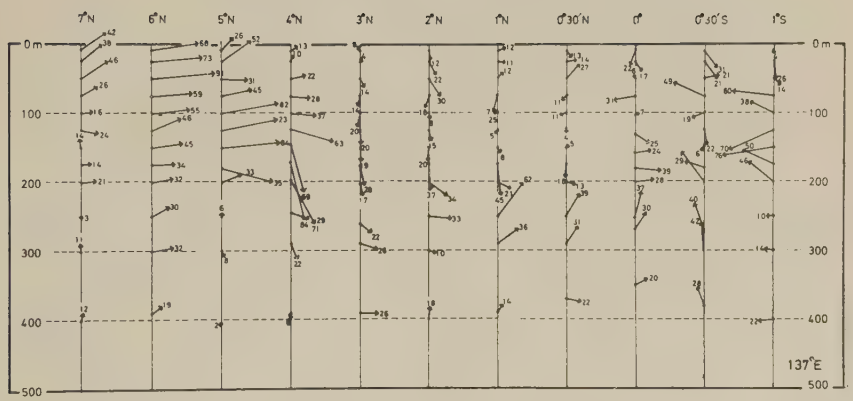


b

Fig. 2. Vertical distributions of velocities in cm/sec measured relative to a depth of 700 m to 900 m at 127°E (a), 130°E (b), 133°30′E (c) and 137°E. (d).


7°N
6°N
5°N
4°N
3°N
2°N
1°N
0°30'N
0°
0 m
100
200
300
400
500
133°30'E

c


7°N
6°N
5°N
4°N
3°N
2°N
1°N
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0°
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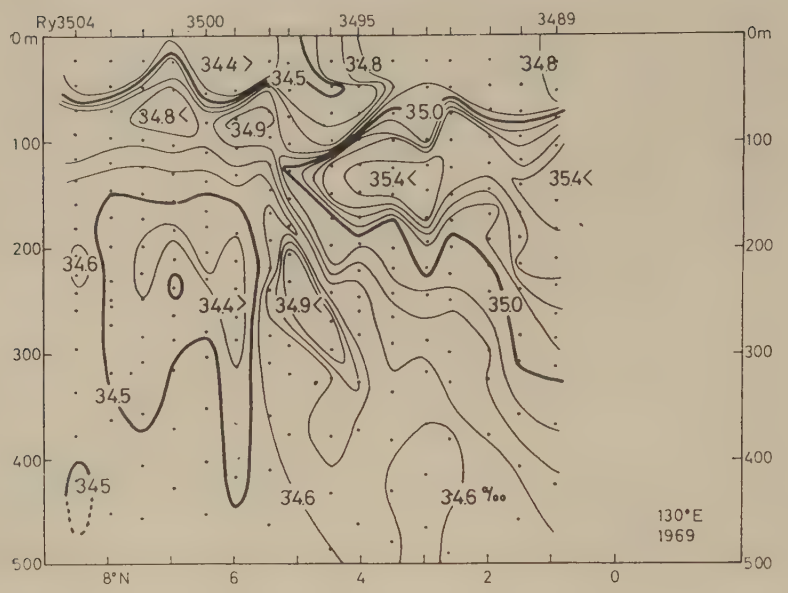
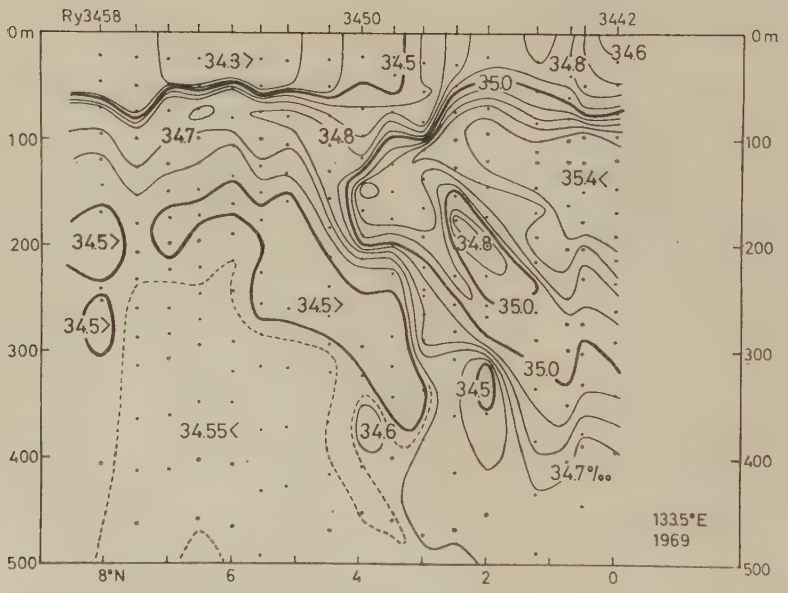
1°S
0 m
100
200
300
400
500
137°E

d

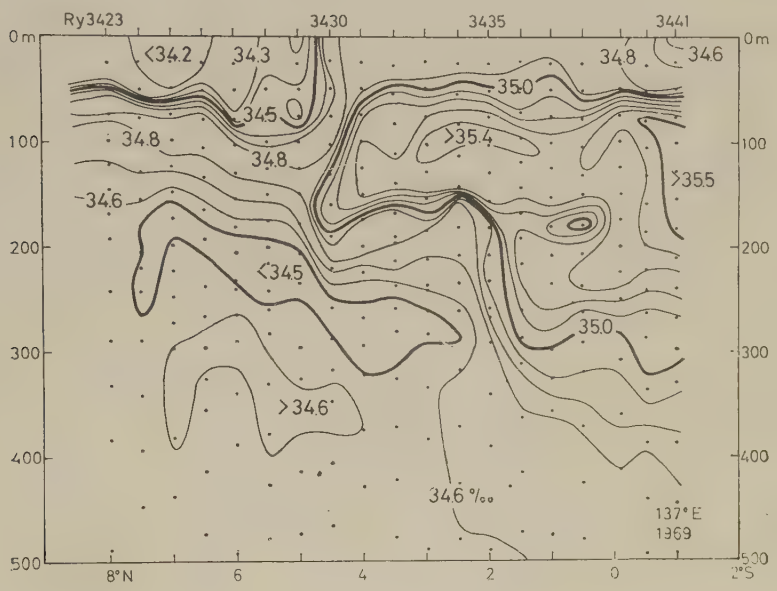


a



b

Fig. 3. The salinity section south of 8°N at 130°E, (a) 133°30′E (b) and 137°E (c).


Ry3423
3430
3435
3441
0m
100
200
300
400
500
<34.2
34.3
34.5
34.8
34.8
34.6
35.0
>35.4
>35.5
<34.5
35.0
>34.6
34.6 ‰
137°E
1969
8°N
6
4
2
0
2°S

c

coast of Mindanao Island. According to these results, the Mindanao Current shows the maximum speed at 150 to 200 m depth, and the fluctuation of speed at the depth is 0.4 kt.

(5) According to the results of current measurement along 17°N, east of Palanan Point, Luzon Island, the current flows northward in two strong current zone, one near the coast and the other near 124°E. The maximum speed in each zone is about 1.8 kt and 1.4 kt at the sea surface respectively, and the northward current extends to the depth of about 600 m, the width of current is 100 miles or so (Fig. 4).

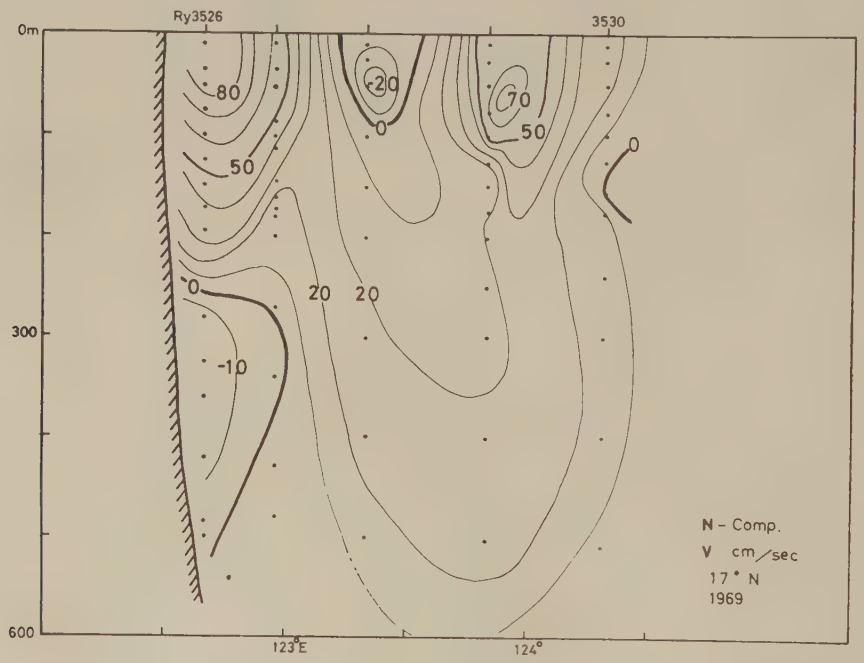


Fig. 4. Vertical distribution of north component of velocity along 17°N, east of Palanan Point.

(6) The vertical distributions of temperature, salinity, thermosteric anomaly and oxygen content at 137°E are shown for demonstrating the gross features of the oceanographic conditions in the surveyed region (Fig. 5).

The distributions of the chemical elements are as follows;

(1) The contents of phosphate phosphorus, silicate silicon and nitrate nitrogen in the vertical section along 137°E are in the ranges of 0—3.3 μg-atoms/*l*, 3—165 μg-atoms/*l* and 0.1—42.5 μg-atoms/*l* respectively. These nutrient salts are

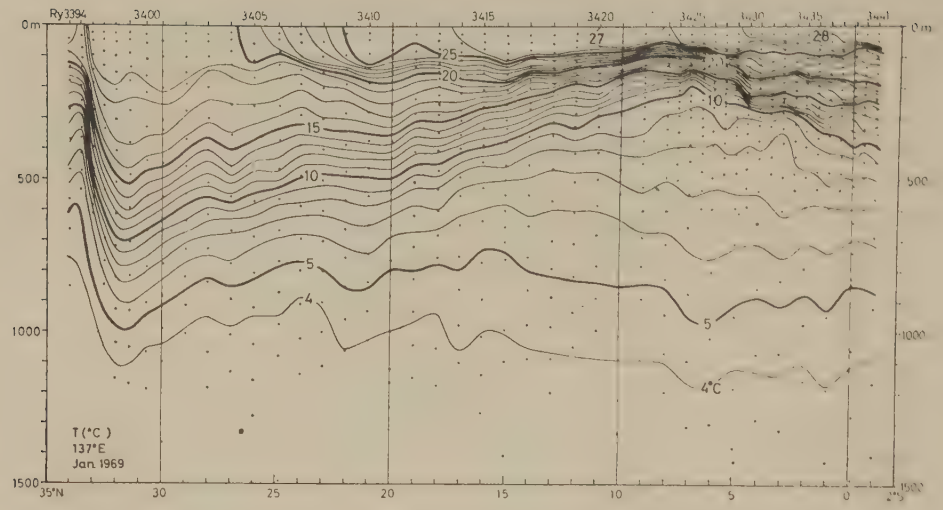
remarkably less in surface water except the region north of the Kuroshio, and increase gradually with depth. The maxima of contents appear at the depth from 1,000 to 1,400 m for phosphate phosphorus and nitrate nitrogen, and at the depth of about 2,500 m for silicate silicon. Below the maxima, the nutrient salts decrease gradually with depth, and rather vary geographically, being lower in the equatorial region and increasing toward the north. The nitrite nitrogen has its maximum content around the depth of 100 m, and below the depth of 300 m the trace amount is observed. The ammonia nitrogen and the highest content (1.4 μg-atmos/l) is observed at 16°N.

(2) The partial pressure of carbon dioxide in the atmosphere along 137°E ranges between 320 and 336 ppm, being low in the low latitudes and increasing toward the north. The partial pressure of carbon dioxide in surface water varies locally in the range between 290 and 397 ppm. It is higher by about 50 ppm in sea water than in the atmosphere in the equatorial region south of 8°N, and is almost in almost equilibrium to the atmospheric partial pressure between 9°N and 16°N, and is lower by about 20 ppm north of 16°N.

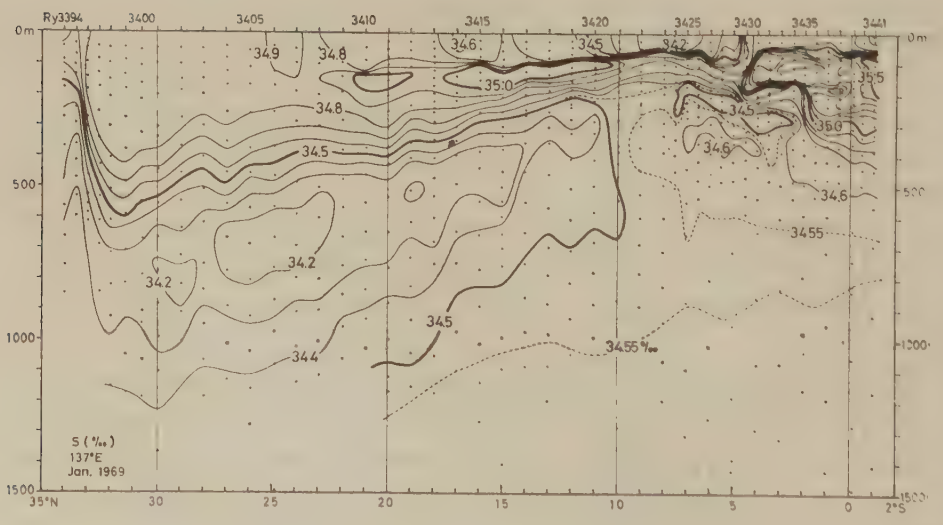
The distributions of plankton are as follows;

(1) The distributions of zooplankton biomasses expressed by means of the displacement volume between the surface and the depth of about 150 m in the winter of 1969 in the western North Pacific are similar to those in the same season of 1968. Along 137°E fairly large biomasses are found in 27°N and 26°N. These facts were also observed in 1968. In the North Equatorial Current the biomass is very small as in 1968. On the contrary, there are large biomasses in the Equatorial Countercurrent and south of the current. The densely populated areas with zooplankton are seen from 8°N to 1°S in the present survey, but they existed between 7°N and 1°S in 1968. Along 133°30′E, 130°E and 127°E, the distributions of biomasses are also similar to those in 137°E. As compared with 1968, large biomasses in 130°E are conspicuous in 1969.

(2) The concentration of chlorophyll, principally chlorophyll *a*, and after acidification phaeophytin, was measured in 90% acetone extracts using a fluorophotometer (HITACHI FPL-2). Chlorophyll concentrations in the section along 137°E are high in the surface layer above 40 m in and north of the Kuroshio, being more than 0.6 μg per liter of water. From the souther edge of the Kuroshio, 32°30′N, to 15°N, the chlorophyll contents are fairly low, less than 0.3 μg. The area with low chlorophyll content is divided into two parts; the northern part having the maximum concentrations of chlorophyll at the surface

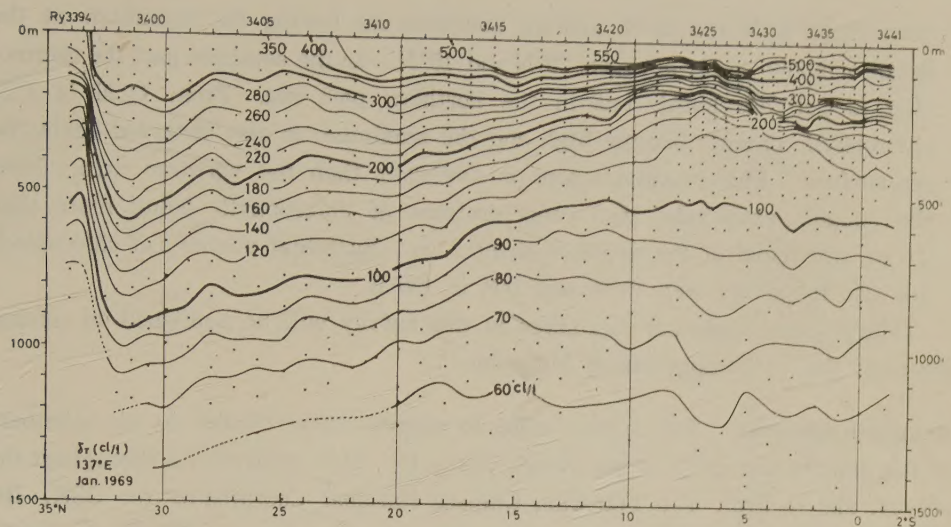


a

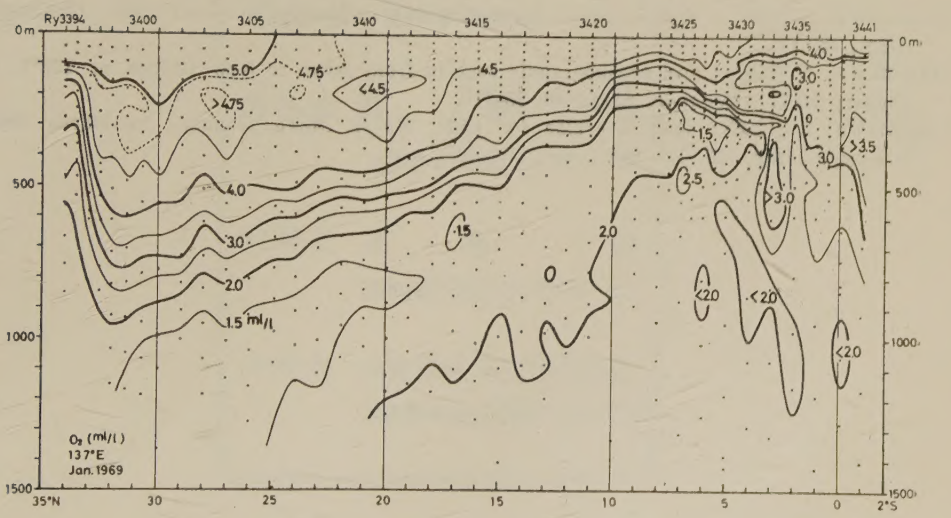


b

Fig. 5. Vertical sections of temperature (a), salinity (b), thermosteric anomaly (c) and oxygen content (d) at 137°E in January 1969.


Ry3394
3400
3405
3410
3415
3420
3425
3430
3435
3441
0m
500
1000
1500
350
400
500
550
280
260
240
220
200
180
160
140
120
100
90
80
70
60 cl/l
300
200
100
δ_T (cl/l)
137°E
Jan. 1969
35°N
30
25
20
15
10
5
0
2°S

c


Ry3394
3400
3405
3410
3415
3420
3425
3430
3435
3441
0m
500
1000
1500
5.0
4.75
>4.75
<4.5
4.5
4.0
3.0
2.0
1.5 ml/l
1.5
2.0
0
<2.0
<2.0
<2.0
4.0
3.0
1.5
0
3.5
>3.5
3.0
O_2 (ml/l)
137°E
Jan. 1969
35°N
30
25
20
15
10
5
0
2°S

d

between 32°30'N and 24°N, and the southern one having the maximum at the depth of about 100 m between 25°N and 15°N. In the southern part the thermocline is well developed, but in the northern part not. From 11°N to 1°S, chlorophyll contents become high and the maximum is also observed below the sea surface. This maximum zone is shallower than the northern area. Near the equator extremely high concentrations of chlorophyll, being more than 2.0 μg, are found at the depth of about 50 m. Such high contents of chlorophyll are also found between 10°N and 1°N in 130°E.

The detailed studies of data taken in this survey will be published on coming issues of the "Oceanographical Magazine".

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